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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

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APRIL 1958

PROPERTIES OF WATER, PART II. THE EQUATION OF STATE OF WATER AND SEA WATER AT LOW TEMPERATURES AND PRESSURES*

CARL ECKART

ABSTRACT. The empirical data on the equation of state of pure water have been reviewed and analyzed. Systematic errors in the isotherms are estimated at $\pm$ 2p x 10^{-7} ml/gm, and the probable error of the best values of specific volume is not less than $\pm$ 2 x 10^{-4} ml/gm, in the range of $0° \leqslant T \leqslant 40°C$ and $0 \leqslant p \leqslant 1000$ atm. In this region and within this error, the data can be represented by the Tumlirz equation

$$(p + p_o)\ (v - v_o) = \lambda,$$

where p = pressure, v = specific volume, while p_o, v_o and λ are empirical functions of temperature.

The quantity v_o seems to be actually independent of the temperature and p_o and λ can be sufficiently well approximated by quadratic functions of temperature. More experimental work is needed to establish these functions with the required precision.

The much less numerous data on sea water are also considered. They are certainly no more precise than the data for pure water, and can also be represneted by the above equation, provided p_o and λ are quadratic functions of temperature and linear functions of salinity; v_o appears to be independent of salinity as well as temperature. The available data do not enable the dependence of the isotherms on salinity to be determined with precision, and more experimental work is needed. Until this becomes available, the formulae derived here must be considered doubtful, but no more so than others in current use.

INTRODUCTION

The primary objectives of this work were (1) a critical examination of the present P-v-T data on sea water, and (2) the construction of a simple but currently adequate empirical formula. The ranges of interest are: pressure, 1 to 1000 atmospheres; temperature, 0 to 40°C; salinity, 0 to 40 gm/kg.

In order to achieve the first objective, it was found necessary to examine the P-v-T data on pure water, partly because the data on sea water are too few to permit an estimate of probable error and partly because the sea water measurements are all relative to distilled water. It has been assumed that the errors in the sea water data are no greater than those of the pure water data. The random errors of the sea water data, insofar as they can be estimated at all, seem to be admirably small; but it is certain that they contain larger systematic errors.

The second objective has not been fully achieved, as will appear below. After the rather lengthy computations had been finally revised, further data and information came to light. It is believed that the formulae would not be significantly changed by a recomputation at this time. The existing sets of data are inconsistent by more than the experimenters' estimates of errors, and

* Contribution from the Scripps Institution of Oceanography, New Series No. 980.

this is reflected by a corresponding uncertainty in the parameters of the equation here derived.

The P-v-T data on water and sea water have been summarized by many empirical formulae, some of them quite elaborate.[1] One of the simplest of these, first proposed by O. Tumlirz (1909),

$$(p + p_o)\ (v - v_o) = \lambda, \tag{1}$$

is entirely adequate for low temperatures and pressures. In Eq. (1), p is the pressure, v the specific volume, and p_o, v_o and λ are functions of the temperature T. Following Dorsey (1940), v will be measured in ml/gm and p will be given in atmospheres = 1.03323 kg/cm² = 1.01325 bar. The quantity

$$\Delta = 10^4(1 - v) \tag{2}$$

is used by Dorsey and will be useful here.[2]

In terms of Δ, Eq. (1) becomes

$$(p - p_o)\ (\Delta_o - \Delta) = 10^4\ \lambda = K. \tag{3}$$

For many purposes, it is convenient to pick a standard pressure, p_1, and its associated Δ_1; then K may be eliminated from Eq. (3):

$$(p - p_1)\ \Delta_o - (\Delta - \Delta_1)\ p_o = (p\Delta - p_1\Delta_1). \tag{4}$$

Defining

$$x = (p - p_1)/(p\Delta - p_1\Delta_1), \tag{5}$$

$$y = (\Delta - \Delta_1)/(p\Delta - p_1\Delta_1), \tag{6}$$

Eq. (4) becomes

$$x\Delta_o - yp_o = 1. \tag{7}$$

The validity of Eq. (1) can therefore be tested by plotting x, y as the coordinates of points, and examining the deviations of these points from a straight line. If Eq. (1) is valid, the intercepts of this line are $1/\Delta_o$ and $-1/p_o$.

PURE WATER

1. *The Data.*—The most convenient tabulation of P-v-T data for pure water is given by Dorsey (1940) as Tables 95-I to 95-IX of his *Properties of Ordinary Water-Substance*. The data of V. W. Ekman (1908) on pure water appear to have escaped Dorsey's attention. A preliminary version of Kennedy's (1957) data was kindly made available to the author. This preliminary version differs from the final version, but for the reasons mentioned above, the present computations have not been correspondingly revised.

Intercomparison of these eleven sets of measurements shows that they can be divided into two groups, those in the first group being mutually consistent within ± 2 units of Δ, and those in the second being inconsistent with each other and with the first group. The first group consists of:

Dorsey's Table 95-I, due to E.-H. Amagat (1893);
Dorsey's Table 95-II, due to E.-H. Amagat;
Kennedy's preliminary data;
Ekman's data.

[1] W. T. Holser will summarize the formulae for pure water in another paper of this series. The standard formulae for sea water contain 37 parameters and are to be found in the book by Bjerknes and Sandström, (1910).

[2] In the oceanographic literature the quantity

$$\sigma = 10^3(1 - v)/v$$

is common, and 1 – v is often denoted by δ or $\Delta\alpha$.

In general, only this first group of measurements will be considered here, except that Dorsey's Table 95-III, due to P. W. Bridgman (1913), will be used for an investigation of high pressures.

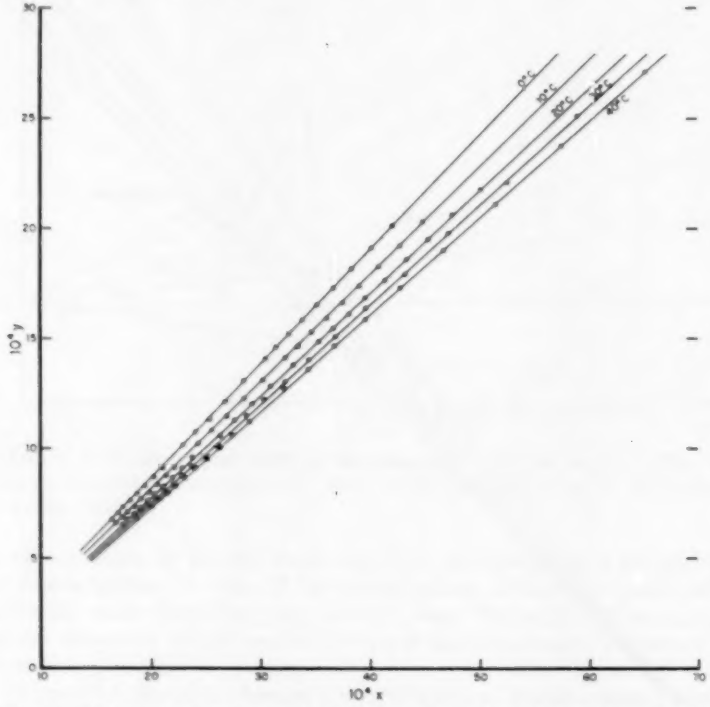

Fig. 1. Isotherms of pure water in the x – y plane: data of E.-H. Amagat (Dorsey, Table 95-I); $p_1 = 500$ atm.

After this paper was already in rough draft, the data of R. E. Gibson and O. H. Loefler (1941) were called to the author's attention by W. T. Holser.

2. *First Test of the Equation.*—Using Dorsey's Table 95-I, values of x and y were calculated using $p_1 = 500$ atm.; they are plotted as figure 1. The pressure range is 1 to 1000 atm., and the 0° to 40°C isotherms are seen to be good straight lines. Bridgman's data, Table 95-III, extend over a larger pressure range, being terminated only by the formation of Ice-VI. The x, y graphs of these data are shown as figure 2. There are systematic departures from the straight lines above $p = 3000$ atm., and there are noticeable irregularities in the spacing of the isotherms. These latter are probably due to experimental error, the former perhaps to inadequacy of the Tumlirz equation (1909).

It is to be concluded that the Tumlirz equation fits the data with some precision for $p \leqslant 3000$ atm., and $0 \leqslant T \leqslant 40°$.

3. *Second Test of the Equation.*—Figures 1 and 2 cannot be drawn on a sufficiently large scale to exhibit the detailed nature of the agreement between

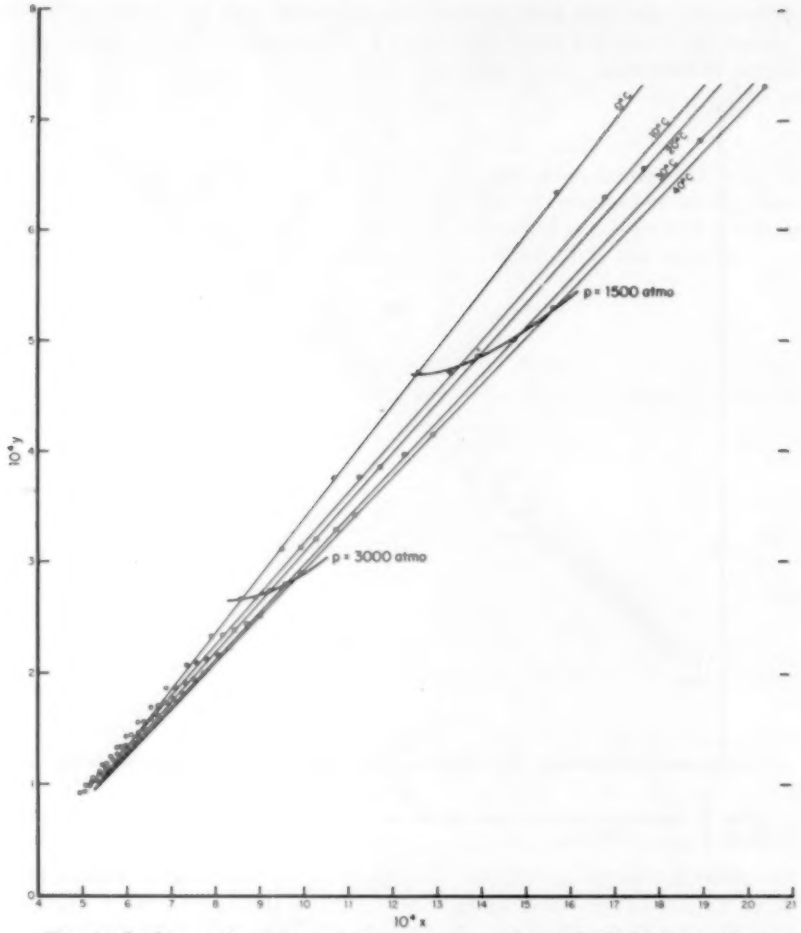

Fig. 2. Isotherms of pure water in the x – y plane: data of P. W. Bridgman (Dorsey, Table 95-III); $p_1 = 500$ atm.

the equation and the data. One would like to sight along the straight lines with a telescope. This can be accomplished mathematically by the transformation

$$\xi = \tfrac{1}{2}y, \qquad \eta = x - 2y. \tag{8}$$

In the ξ, η plane, the Tumlirz equation becomes

$$\alpha\xi + \beta\eta = 1, \tag{9}$$

$$\alpha = 4\Delta_o - 2p_o, \qquad \beta = \Delta_o. \tag{10}$$

Figure 3 shows the same data as figure 1, replotted in this fashion. It is seen that this provides a more searching test of the relation between the data

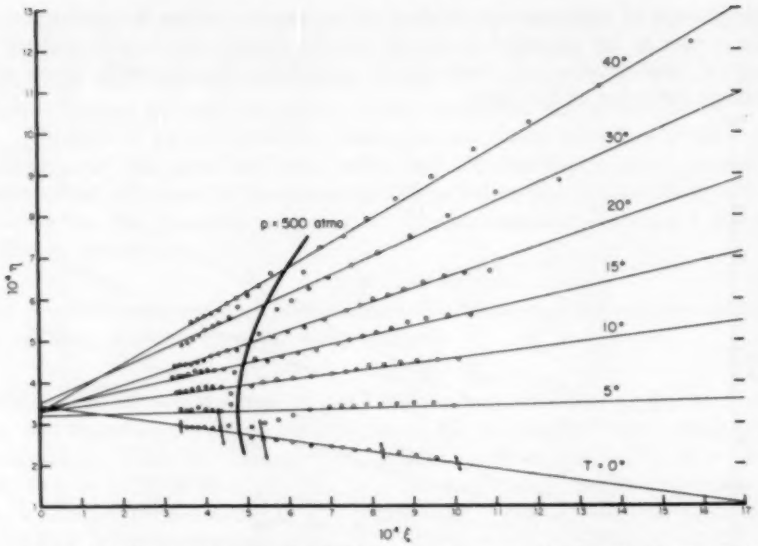

Fig. 3. Isotherms of pure water in the $\xi - \eta$ plane: same data as fig. 1. The double arrows indicate the displacements that would result from changes of $\pm$ 10^{-4} ml/gm in the specific volume.

and the equation. Systematic departures from the straight lines are apparent. The double arrows on some of the plotted points indicate the displacements that would result from changing Δ by $\pm$ 1 unit. The lengths of these arrows, and the sensitivity of the method, increase as the standard pressure $p_1 =$ 500 atm. is approached.

Figure 4 is based on Amagat's second series of measurements (Dorsey's Table 95-II). This, like the preceding figure, shows systematic departures from the straight lines—but the systems are not the same in the two cases. Both Tables 95-I and 95-II give values of Δ to 0.1 unit; one concludes, however, that there are systematic errors of at least one unit.

Figure 5 shows Kennedy's preliminary data, plotted in the same way. While Amagat smoothed his data to $\pm$ 0.1 unit of Δ while interpolating, Kennedy estimated his error at about $\pm$ 1 unit, and smoothed accordingly.

Amagat's first set of data, figure 3, yields a $\xi - \eta$ diagram that shows smaller departures of the isotherms from straight lines than do the two other sets. This might indicate that Amagat's first set is the most reliable of the three. Without more information about Amagat's method of interpolation, the writer is disinclined to adopt this view. It seems more reasonable to suppose that the differences among the three figures disclose the presence of systematic errors in all of them. It should be noted, specifically, that compatibility with the Tumlirz equation is no guarantee of the absence of such errors. More generally, experience in other fields indicates that internal consistency and smoothness do not necessarily imply lack of error, and that, while the inter-

comparison of independent published measurements enables the magnitude of the errors to be estimated, it rarely permits them to be located without a critical consideration of experimental procedures. Such considerations are outside the scope of this paper.

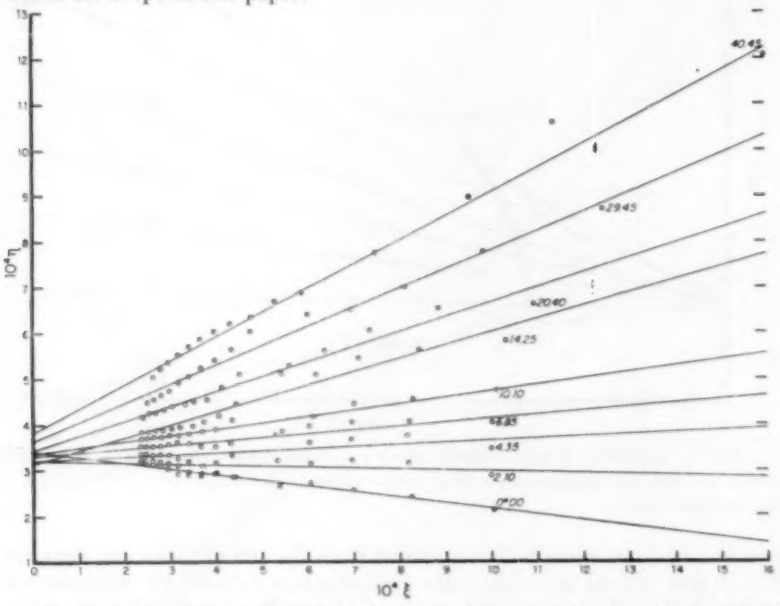

Fig. 4. Isotherms of pure water in the $\xi - \eta$ plane: second series of data of E.-H. Amagat (Dorsey, Table 95-II).

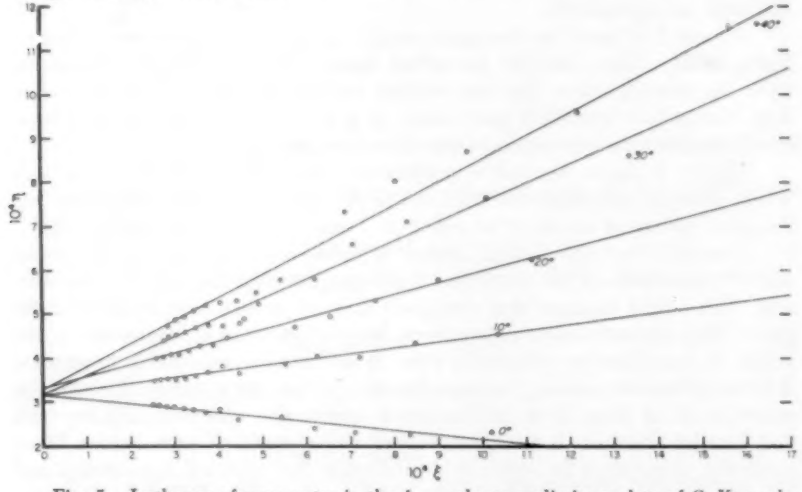

Fig. 5. Isotherms of pure water in the $\xi - \eta$ plane: preliminary data of G. Kennedy.

One may reach the provisional conclusion: the precision of the specific volume data is not better than ± 2 units of Δ, and the Tumlirz equation represents the isotherms adequately in the low pressure and temperature region. Further evidence in support of this conclusion will appear below.

4. *The Data of Ekman (1908).*—These were not found until most of the calculations for this paper had been made. They will therefore be given separate treatment at all stages of the discussion. Ekman does not tabulate the specific volume but the "apparent compression". The compression (between 1 and p atm.) is defined as

$$k = 1 - v(p, T)/v(1, T), \tag{11}$$

the apparent compression k′ differs from this because of the compressibility of the glass in which the water was contained:

$$k' = (k - k_g)/(1 - k_g), \tag{12}$$

where k_g is the compression of glass. Ekman was uncertain of the value of k_g, and therefore published only values of k′; for distilled water, these are given in his Table 10 (Ekman, 1908, p. 32). He first suggests that $k_g = pv$, where $v = 0.225 \times 10^{-6}$ per bar; in a footnote added after correspondence with Amagat, he revises this to $v = 0.215 \times 10^{-6}$. This second value has been used here in calculating k from k′.

It is also necessary to know v(1, T) in order to obtain v(p, T): these data were taken from Dorsey's Table 94-III (revised Chappuis Table, 1937). The resulting values of Δ are given as table 1.

Ekman had no direct means of measuring pressure to the required precision. Consequently, he made indirect measurements, using the compression of distilled water at 0°C for this purpose. In order to convert these to pressures, he used a mean of Amagat's two 0° isotherms. It has been seen that these differ by several tenths of a unit in Δ, and probably contain even larger systematic errors. Ekman's values of the specific volume at T = 0°C are therefore necessarily an average of those of Amagat.

At other temperatures, they differ from those of Amagat by as much as 3 units of Δ. It is impossible to determine how the error thus disclosed is divided between the Amagat and Ekman values. Table 1 shows that Ekman's repetitions always agree to better than 0.1 unit, and often coincide to 0.01. This would lead one to give them the greater weight, were it not for their forced agreement with Amagat's 0°C isotherm: Amagat's errors influence Ekman's "measured" values in a systematic manner.

Even if the Ekman data were more accurate than the Amagat data, it would not be possible to use them for a searching test of the Tumlirz equation. Each of Ekman's isotherms is defined by just two measurements at each of three pressures: consequently the three parameters of Tumlirz' equation can always be chosen to yield the exact average of the two values of Δ at each pressure. At best, one can assume the Tumlirz equation and calculate these parameters.

5. *Temperature Dependence of the Isotherms.*—The straight lines shown on figures 1 to 5 were determined by least squares. Each yields a determination of

TABLE 1

Values of $\Delta = 10^4 (1 - v)$ deduced from Ekman's (1908, p. 32.) Measurements on Distilled Water

T / p atm.	4° .97	9° .97	14° .96	19° .96
203.7	95.32 95.32	90.08 90.11	82.10 82.16	71.66 71.64
408.9	185.55 185.62	178.04 178.04	168.35 168.35	156.46 156.40
607.4	267.44 267.44	258.01 258.03	246.77 246.80	233.69 233.67

the parameters α and β in Eq. (9), for the corresponding temperatures. These various determinations are plotted on figure 6, a and b.

The systematic variation of β is very slight, if not negligible. The mean of the Kennedy and Amagat values is $\beta = 3020$, with a standard deviation of 158. The straight line $\beta = 3088 - 3.930T$ fits these same data with a standard deviation of 146. One is therefore justified in concluding that β is independent of temperature, and has the value

$$\beta = 3020 = \Delta_0. \tag{13}$$

The four values from the Ekman data yield $\beta = 2855 \pm 15$ and are thus barely consistent with the others. They do confirm the constancy of β, and

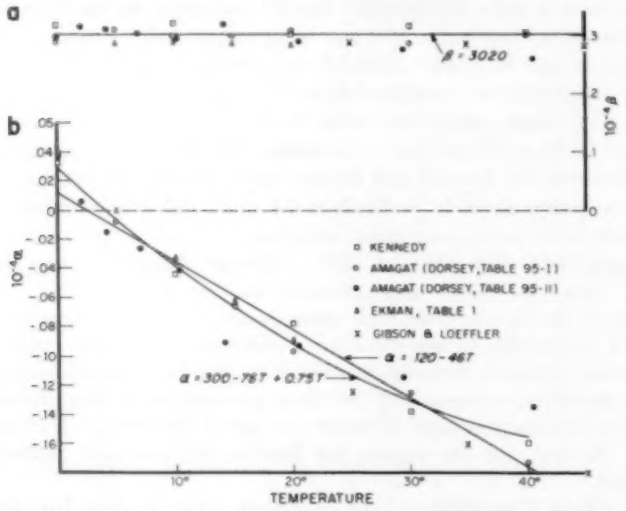

Fig. 6. Values of α and β (Eq. 9) for various temperatures and various experimenters.

would have somewhat lowered the overall average. Values of β were calculated from the Gibson and Loeffler data for T = 25°, 35°, 45° and 55°; while these data extend to 85°, it seemed undesirable to include the higher temperatures in the present study. The four selected values yielded β = 2841 ± 70. The value of β = 3020 has been used in the following calculations, for the reasons explained above.

The values of α show a definite trend, and are represented by the linear formula

$$\alpha = 120 - 46T \tag{14}$$

with a standard deviation of 168. This is shown by the straight line on figure 6b; the values at T = 0° and 40° deviate from it by undesireably large amounts. The quadratic formula

$$\alpha = 300 - 76T + 0.75T^2 \tag{15}$$

represents the data with a standard deviation of 110. The reduction from 168 to 110 is scarcely large enough to justify Eq. (15), but the desire to reduce the deviation at T = 0°C leads to its provisional adoption.

The inclusion of the Ekman and Gibson and Loeffler values of α would have reduced the statistical significance of the quadratic term still further, and would have increased the coefficient 46 in Eq. (14). In particular, it is noted that the Ekman values trend toward Amagat's points at 0°C; in view of Ekman's method of measuring pressure, this must be somewhat discounted.

Again, Eq. (15) has been used in the following calculations. This is even less justifiable than the corresponding decision regarding Eq. (13), since it

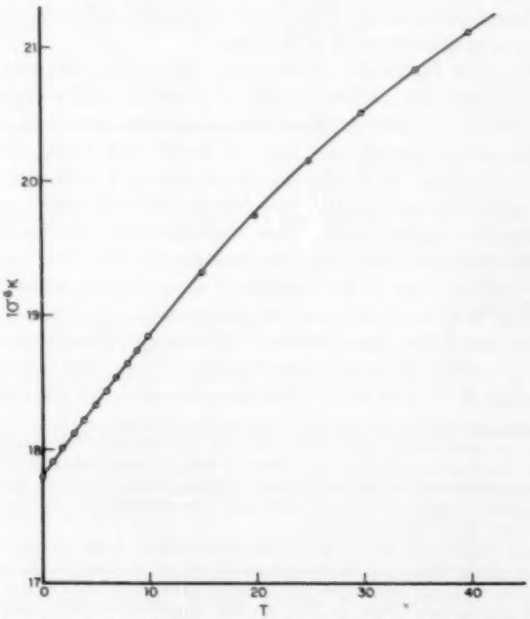

Fig. 7. Variation of K with temperature: data of Chappuis (Dorsey, Table 94-III).

gives undue weight to the three values of α at 0°C. It would be very desireable to have more measurements in the range $0° \leqslant T \leqslant 15°C$. Perhaps the use of Eq. (15) can be condoned until such measurements become available.

Since $\alpha = 4\Delta_o - 2p_o$, Eq. (13) and (15) imply

$$p_o = 5890 + 38T - 0.375T^2. \tag{16}$$

It remains to find the dependence of K, Eq. (3), on the temperature. Having shown that Eq. (3) represents the isotherms to an adequate precision, and having determined p_o and Δ_o, it suffices to consider only a single value of p and its associated value of Δ at each temperature. Since the thermal expansion at $p = 1$ atmosphere has been determined with great precision, this pressure is the logical choice. Dorsey's Table 94-III provides values of the specific volume at $p = 1$, which are probably accurate to $\pm$ 0.05 units of Δ, a greater precision than is available at any other pressure.[3] The values of K determined by substituting from this table into Eq. (3) are plotted as figure 7. They are fitted by the quadratic equation

$$10^{-6}K = 17.795 + .1125T - .000745T^2 \tag{17}$$

with a standard deviation of 0.003. The quadratic term in Eq. (17) is certainly significant.

These methods of determining the parameters Δ_o, p_o and K as functions of T are not the best. It would have been preferable to apply least squares methods directly to Eq. (3). The above methods, based on values of x and y, are not only more laborious, but give undue weight to measurements near the arbitrarily chosen pressure p_1. They were originally adopted because they are well suited for a graphical study of the data.

6. *Final Test of the Equation.*—Using Eqs. (3), (13), (16) and (17), values of Δ were calculated for various values of p and T, and compared with the tabulated values of the first group of measurements described above. As examples of the second group, the data of Smith and Keyes, from Dorsey's Table 95-VI, and those of Bridgman, from Dorsey's Table 95-III, were included. The differences are plotted on figure 8,[4] the ordinates being $\Delta - \Delta_{calc.}$. It is seen that, for a given temperature and observer, these show systematic variations with pressure. However, the systems are different for different experimenters, and are thus to be considered as systematic errors in the measurements, rather than as a failure of the equations.

It is to be noted that the systematic differences among experimenters are greatest at $T = 40°C$. The systematic departures from the equations adopted here are greatest at 0° and 40°C. This has attracted the attention of several

[3] Since all measurements of v at elevated pressures are relative to v(1, T) a similar choice must be made in constructing any tables of v(p, T). Dorsey has apparently used his Table 94-III in constructing most of his tables of v(p, T), but Table 95-VI, due to Smith and Keyes, is an exception. In the preliminary version of the Kennedy data studied here, the Smith and Keyes values of v(1, T) were used by mistake; in the published version of these data, Table 94-III has been used.

[4] In figure 8 the Smith and Keyes points deviate strongly from all the others. This is because their tabulated values are derived from an empricial equation that is extrapolated below 30°C, a fact which is not apparent in Dorsey's retabulation. The only experimental values of Smith and Keyes (1934, p. 297) in this range are at 30°C, and they are generally within the range of values of other investigators.

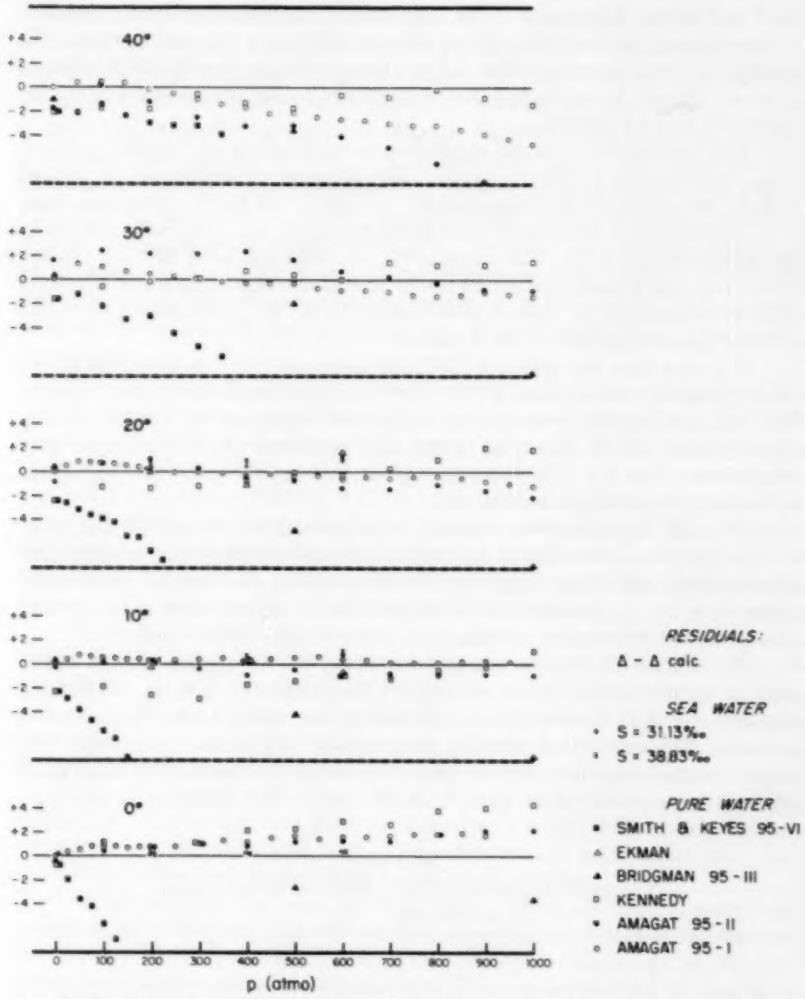

Fig. 8. Differences between observed and calculated values of $\Delta = 10^{-4}(1-v)$: various observers; Eqs. (3), (13), (16) and (17).

readers of the manuscript, and has been checked to make certain that it is not due to any arithmetic mistakes.

The systematic errors thus disclosed affect the slope of the isotherms, and may be estimated at about $\pm$ 2p x 10^{-7} ml/gm. In adition, there are more or less random errors in the data, which may be estimated at $\pm 2 \times 10^{-4}$ ml/gm. There are also errors in the spacing of the isotherms, which are not disclosed by figure 8, but are treated below. These estimates of error are not very certain, and are likely to prove optimistic.

7. *The Thermal Expansion.*—The isothermal compressibility, being essentially the slope of the isotherms, is not influenced by errors in their spacing. The systematic pressure errors affect it in a manner that is roughly independent of pressure. It may be estimated that these errors in the compressibility are between 0.5 and 1.0 percent.

The coefficient of thermal expansion is essentially $a = (\partial v/\partial T)p$, and is thus very sensitive to spacing errors. The existence of such errors is implied both by the errors in the compressibility and in p_o and K. To investigate them, a was calculated in two ways: once using the linear Eq. (14) and once using the quadratic Eq. (15). The value of $\Delta_o = 3020$ was used in both calculations, and interpolation of the values of K was avoided. This made it certain that the values of a at 1 atmosphere would be the accepted ones. The results of these calculations are given in table 2.

It is seen that the values at 20°C are uncertain by 3 to 5 percent at elevated pressures, while those at T = 0°C are much more seriously in doubt. This can perhaps be remedied by additional experimental work at closely spaced values of T. However, unless the precision of the compressibility measurement can be increased, not all of the error in the spacing of the isotherms will be eliminated by this.

It would be desirable to develop experimental techniques for the more or less direct measurement of a at elevated pressures. It is also possible that other methods of fitting algebraic formulae to the data would yield better values of a, but it is very doubtful whether much improvement of the present values can be achieved by modified calculational procedures alone.

The coefficient of thermal expansion is needed for the calculation of the ratio of specific heats, γ. It is known $\gamma - 1$ is proportional to a^2, so that this quantity will also be seriously in doubt at low temperatures and elevated pressures. The velocity of sound is proportional to $\gamma^{1/2}$; since a is small, the proportionate error in γ will be small. However, the adiabatic temperature gradient is proportional to $(\gamma - 1)/a$, i.e., to a. This quantity is therefore seriously in doubt.

TABLE 2

Thermal expansion, $10^5 a$, calculated in two ways
(a in $cm^3\ gm^{-1}\ deg^{-1}$)

p atm.	T = 0°		T = 20°	
	Linear Eq. (14)	Quadratic Eq. (15)	Linear Eq. (14)	Quadratic Eq. (15)
1	–6	–6	21	21
100	–4	–3	22	23
500	+3	+8	26	26
1000	9	19	31	30

SEA WATER

1. *The Data.*—The specific volume of sea water at elevated pressures has been determined only once, by V. W. Ekman (1908). The measurements were

all made on two halves of a single sample of sea water, one half having been diluted with distilled water, the other concentrated by evaporation. No investigation of the validity of this procedure was made. The original publication contains only values of the apparent compressibility (see above). Values of the specific volume calculated by Ekman "as directly as possible from the compressibilities measured, using the necessary additional data from Amagat and Martin Knudsen" were published[5] as Table C in Bjerknes and Sandström (1910). The need for Amagat's data has been indicated above. Knudsen (1902) measured the thermal expansion of sea water of various salinities at atmospheric pressure; Bjerknes and Sandström, in their Table B, converted this data "from the relative measure employed in the course of the experiments to the values (of specific volume) in absolute units"—i.e., presumably to cm^3/gm.

Dorsey's Table 108 gives the density of sea water at one atmosphere, for various temperatures and salinities; he comments that this table is "derived from a table by J. A. Beattie (1928) based primarily on the data given by M. H. C. Knudsen (1902) and J. Y. Buchanan (1875) but with a consideration" of a voluminous literature. These two tables have been compared in a cursory fashion: they seem to agree to about one half unit of Δ. It may therefore be supposed that the use of Bjerknes and Sandström's Table B will not introduce errors as great as those resulting from Ekman's reliance on Amagat's 0°C isotherm for the calculation of his pressures.

Bjerknes and Sandström in their Table C give values of v for p approximately 1, 200, 400 and 600 atmosphere, various values of T, and two values of the salinity: S = 31.13 and 38.53 gm/kg. The values of v are given to 0.01 unit of Δ: Dorsey (1940) copies this table as his Table 109, but rounds the figures to 0.1 unit of Δ. According to the above considerations, the accuracy of these data can hardly be greater than the accuracy of those for pure water.

Current oceanographic conventions require that densities and specific volume be calculated (from field measurements) to 0.1 unit of Δ; they are sometimes calculated to 0.01 unit. These conventions are not realistic, and should be reexamined, for necessity as well as for realism.

2. *The Interpolation Equation for Sea Water.*—There are thus, at most, four points on each isotherm for each of the two salinities. A test of the Tumlirz equation is barely possible by the above procedure, since one of the four points would be sacrificed as the standard. This can be avoided by averaging Eq. (4) over p_1 Δ_1, obtaining

$$(p - \overline{p})\ \Delta_o - (\Delta - \overline{\Delta})\ p_o = (p\Delta - \overline{p\Delta}). \tag{18}$$

If the revised definitions

$$x = (p - \overline{p})/(p\Delta - \overline{p\Delta}), \tag{19}$$

$$y = (\Delta - \overline{\Delta}/(p\Delta - \overline{p\Delta}), \tag{20}$$

are used, each of the straight lines is defined by four points, and yields a

[5] Dorsey copies this table as Table 109 of his book; however, the pressures are in error: they should be increased by 1. atmosphere. The error arises from the oceanographic convention that p denotes "sea pressure," caused by overlying water, exclusive of the atmospheric pressure acting on the sea surface.

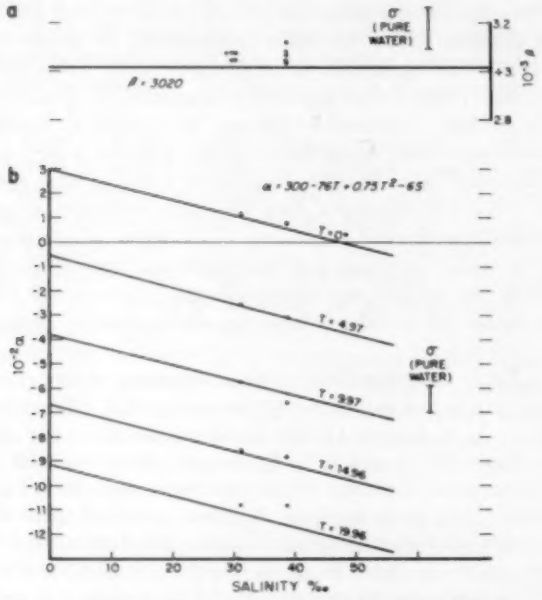

Fig. 9. Values of α and β, Eq. (9), for sea water of two salinities.

slightly less meager test of the equation. While not conclusive, this test is not inconsistent with the use of the Tumlirz equation for sea water.

Values of α and β were determined by least squares from the values of ξ, η so obtained. These are plotted as figure 9. The values of β, for all values of the temperature and the two values of salinity, all lie within one standard deviation of the value for pure water. One therefore concludes that $\beta = \Delta_0$ is independent of salinity as well as independent of temperature; obviously this conclusion should not be extrapolated to higher salinities without further study.

The values of α show a systematic downward trend with salinity. The total decrease is not more than three standard deviations, but recurs for each temperature. It may therefore be concluded that

$$\alpha = 300 - 76T + 0.75T^2 - 6S, \tag{21}$$

$$p_0 = 5890 + 38T - 0.375T^2 + 3S. \tag{22}$$

The agreement between this formula and the empirical values of α is shown on figure 9b.

Finally, by using the values of v at atmospheric pressure, from Bjerknes and Sandström, Table B, values of K were computed. To a sufficient approximation, these are given by

$$10^{-6}K = 17.795 + 0.1125T - 0.000745T^2 - (0.0380 + 0.0001T)S. \tag{23}$$

The ultimate test of these formulae is, of course, the study of $\Delta - \Delta_{calc.}$ for sea water. These values have been included on figure 8. It is seen that

they are of the same order of magnitude as those for pure water. Possibly one may conclude from this that the specific volume of sea water is known to the same precision as that of pure water at 0°C. Certainly Ekman's experimental work is of high quality, and is internally consistent within narrow limits. However, his measurements were made at very few points, and have not been repeated by others. Under these conditions, the presence of unsuspected systematic errors cannot be excluded, and Eq. (22) and (23) should be used cautiously. They are uncertain, but no more so than the much more elaborate equations currently in use.

CONCLUSIONS

1. The isotherms of pure water are adequately represented by Eq. (1) in the temperature range 0° to 40°C and pressure range 0 to 1000 atm.
2. In this range, v_o is a constant, independent of temperature T, insofar as this can be determined by a statistical examination of the data.
3. In this range, p_o and λ are quadratics in T; the statistical significance of the quadratic term in p_o is very low, however.
4. Additional experimental work, at closely spaced values of T, will be required to determine the variation of p_o more precisely.
5. The random errors in the experimental specific volume determinations are not less than $\pm$ 2 x 10^{-4} ml/gm; systematic errors, increasing with pressure, are not less than $\pm$ 2p x 10^{-7} ml/gm.
6. The empirical isotherms of sea water are adequately represented by Eq. (1), in the same range of temperature and pressure as above, and in the salinity range 0 to 40 gm/kg.
7. In this range, v_o is independent of salinity as well as of temperature.
8. In this range, p_o and λ are linear functions of S, insofar as this dependence can be deduced from the data. The statistical significance of the variation of p_o is not great.
9. Additional experimental work is needed to establish the dependence of p_o and λ on salinity with adequate precision.
10. The errors in the experimental specific volume determinations of sea water, both random and systematic, are not less than in the corresponding measurements for pure water, and the systematic errors may be greater.
11. The systematic errors in the P-v-T data render the isothermal compressibility of water and sea water uncertain by 0.5 to 1.0 percent. Their effect on the coefficient of thermal expansion is considerably greater, especially at low temperatures and high pressures; e.g.: at S = 0, 0°C and 1000 atm., the thermal expansion may be 9 or 19 x 10^{-5} per degree, depending upon the interpretation of the data.
12. Current oceanographic conventions require that densities and specific volumes be calculated (from field measurements) to 0.1 unit of Δ; they are sometimes calculated to 0.01 unit. These conventions are not realistic, and should be reexamined, for necessity as well as for realism.

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INTERACTION OF BASEMENT AND MANTLE DURING FOLDING NEAR RUTLAND, VERMONT

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ABSTRACT. In the Green Mountains near Rutland, Vermont, Precambrian gneiss, schist, quartzite, marble and amphibolite are unconformable beneath Cambro-Ordovician carbonate rocks, phyllite, quartzite and volcanic rocks. During the sharp Paleozoic folding of the surface of angular unconformity, relatively incompetent beds in the basement complex were bent nearly parallel with the unconformity, as revealed by regional mapping of the basement complex. In the absence of bending, basement rocks were sheared parallel to the unconformity. This bending or shearing readjustment is limited to a zone several thousand feet thick just beneath the unconformity.

Relatively competent material in the Precambrian such as massive quartzite has caused folding of Paleozoic rocks on axes parallel to contacts of the quartzite, even though these axes are perpendicular to the normal fold trend of the Green Mountains. This refraction of younger folding was locally accentuated by constriction of the flow of younger material by unusually resistant quartzite buttresses.

Paleozoic rocks had acquired some rigidity at the time of deformation, but the guiding influence during arching of the unconformity was difference in basement competence. These conclusions are compared with observations made in western Norway, the western Alps and the western United States.

INTRODUCTION

The Green Mountains of central and southern Vermont are in part the expression of a great anticline of crystalline Precambrian rocks mantled by a variety of lower Paleozoic strata. The trend of this anticline can be followed from the Canadian border at least as far as Virginia. While mapping in central Vermont, the writer had the opportunity of examining a portion of this anticline in some detail.[1] At that time very little was known of the relations of the gneisses, schists and marbles which on the basis of mineralogic dissimilarity and structural discordance with Paleozoic rocks had been consigned to the Mt. Holly complex (Osberg, 1952, p. 21) of Precambrian age. Although structural interpretation of the Mt. Holly complex is complicated by the unknown origin of many of the Precambrian rocks and the several periods of metamorphism, the geometry of the core rocks was decipherable, as it proved, with rather interesting results.

Highly folded unconformities are certainly not unusual, even those which separate rocks of greatly different age and metamorphic history. It is easy to imagine several ways in which structurally discordant rocks above and below unconformities might interact in the course of folding; it was in fact just this question which prompted closer examination of the anticline of the Green Mountains.[2] In this paper, following a brief review of the geology of the region, this interaction is examined.

[1] Some of the descriptive material of this paper also appears in Brace, 1953, which is a rather detailed study of the stratigraphy, structure and metamorphism of the Rutland area, Vermont. Renewed discussion of the problem seems warranted following observations made in the Alps in 1954 and interesting recent work of Hudson (1955).

[2] Professor J. B. Thompson, Jr., of Harvard University first called my attention to problems of this sort in central Vermont, and throughout the work was a constant source of ideas, encouragement and helpful criticism. He and Professor M. P. Billings reviewed the manuscript.

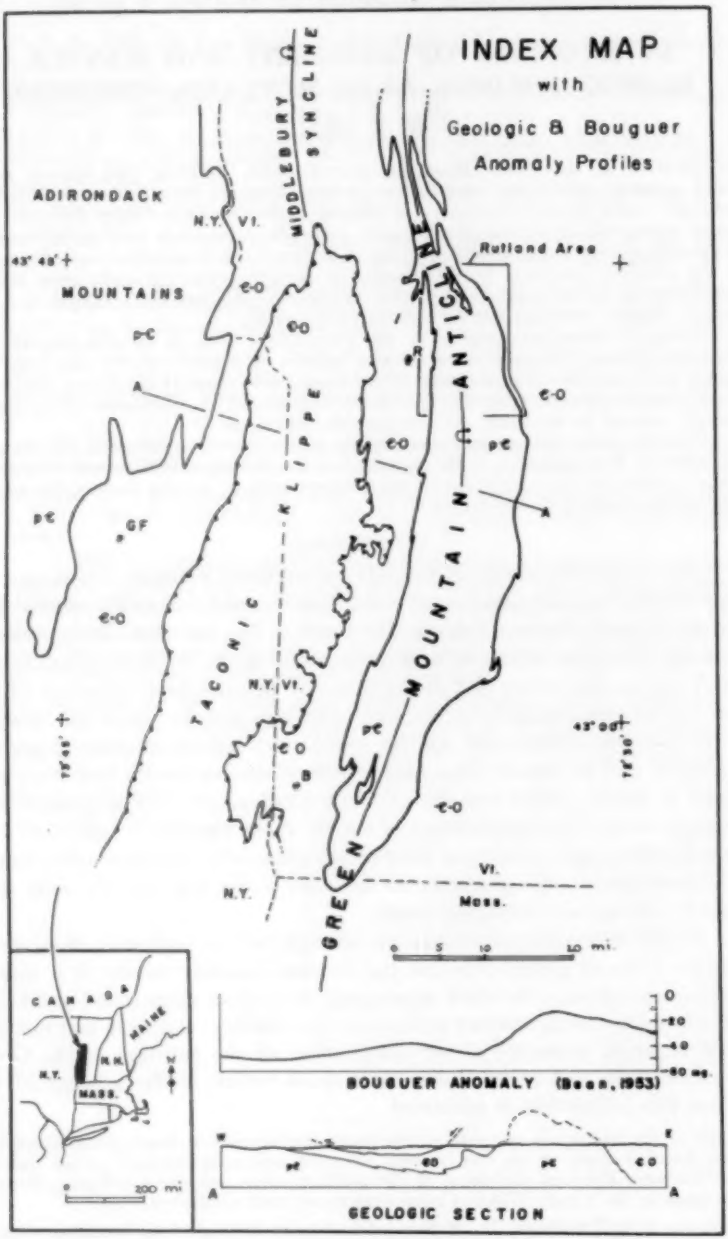

Fig. 1. Index map with geologic and Bouguer anomaly profiles, adapted from Billings et al., 1952 and Bean, 1953. R is Rutland, Vermont, GF is Glen Falls, N. Y., and B is Bennington, Vermont. Cambro-Ordovician (€O) mantling strata rest unconformably on the Precambrian (p€) of the core of the Green Mountain anticline and of the Adirondack mountains. The location of the Rutland area is shown; for detail see figure 2.

Geologic Setting.—The Rutland area, which was studied in detail, includes a portion of the Green Mountain Precambrian basement complex and representative mantling Paleozoic strata (figs. 1, 2). The geology of Rutland and neighboring areas is sketched below; for more complete treatment of regional problems see the references cited.[3] Unfortunately there is very little other detailed work on Green Mountain Precambrian rocks.[4] No attempt will be made to apply conclusions of this study more than incidentally to such issues as the Taconic problem, or the correlation of lower Paleozoic rocks in northern New England.

Throughout nearly the entire length of Vermont the Green Mountain anticline separates dolomite, limestone, quartzite and conglomerate to the west from argillaceous and volcanic rocks to the east (figs. 1, 2). The rocks west of the anticline contain Cambrian or Ordovician fossils whereas those to the east are unfossiliferous. Present correlation of eastern and western sequences is based on stratigraphic position and metamorphic history, for these sequences have not been found in direct contact in Vermont.

West of the Green Mountain anticline, Cambro-Ordovician carbonates and quartzites extend as far as the Adirondack Mountains in which Precambrian rocks of the Grenville appear. (Engel and Engel, 1953). East of the Green Mountain anticline, the contrasting argillaceous and volcanic sequence extends in more or less regular fashion to New Hampshire.

Regional metamorphism in the Rutland area has produced chlorite and biotite in argillaceous rocks west of the Green Mountain anticline; formation of garnet in the eastern sequence indicates a slight increase in grade of metamorphism; metamorphism and deformation are not well dated, but probably occurred in the middle Paleozoic.

The Precambrian rocks in the core of the anticline are divided into the Saltash and Wilcox formations (Brace, 1953, p. 19) and the Mount Holly complex. The Saltash and Wilcox are believed to be younger inasmuch as the grade of metamorphism is lower and they rest unconformably on the Mount Holly. The Wilcox consists of about 3,000 feet of green, white and black schist, which contains thin coarse-grained dolomite beds, and about 500 feet of pegmatitic quartzose gneiss near the base of the formation. The Wilcox formation is only found in a tightly folded syncline in the southeastern part of the Rutland area (fig. 2).

The Saltash formation contains about 1800 feet of grey vitreous quartzite, graphitic phyllite with thin fine-grained dolomite and limestone beds and, at the base of the formation, quartz-albite-sericite-chlorite graywacke. Saltash mineralogy and metamorphism is very similar to that of Paleozoic rocks, but it is separated from both Paleozoic and Mount Holly by angular unconformities.

The Mount Holly contains at least 7,000 feet of the following rock types, in decreasing order of abundance: gneiss, quartzite and quartz schist, amphibilite and greenstone, marble and lime silicate rocks, pegmatite and quartz-

[3] Note particularly Cady, 1945; Billings et al., 1952; Bean, 1953; Zen, (in preparation).

[4] Published work includes Osberg, 1952; MacFayden, 1956; Hawkes, 1941. Valuable unpublished work is Thompson, 1950; Hawkes, 1940; Rosenfeld, 1954.

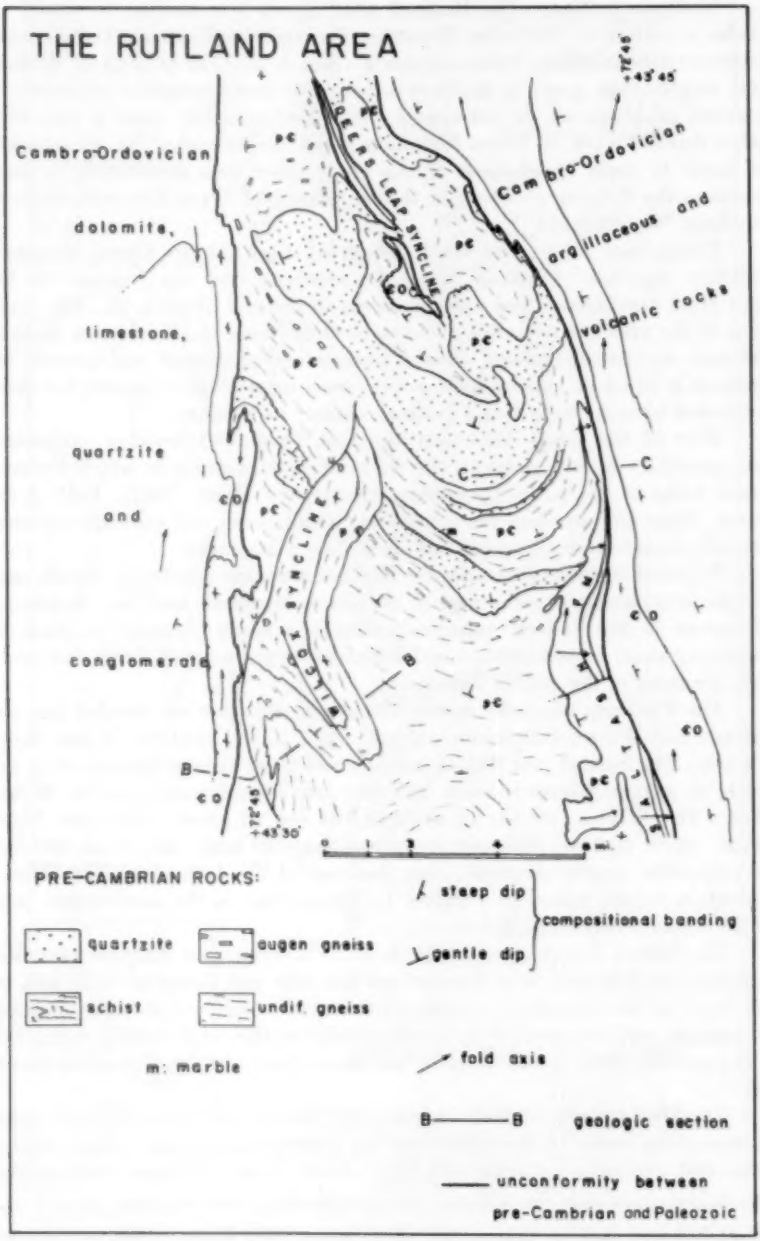
THE RUTLAND AREA
Cambro-Ordovician
dolomite,
limestone,
quartzite
and
conglomerate
Cambro-Ordovician
argillaceous and
volcanic rocks
QUEERS LEAP SYNCLINE
BLEDGE SYNCLINE
pC
C
B
+72°45'
+43°45'
+72°45'
+43°30'
0 1 2 3 4 5 mi.
PRE-CAMBRIAN ROCKS:
quartzite
augen gneiss
schist
undif. gneiss
m: marble
steep dip
gentle dip
compositional banding
fold axis
B——B geologic section
unconformity between
pre-Cambrian and Paleozoic

Fig. 2.

tourmaline veins. The Mount Holly resembles petrographically rocks found in the Adirondack Mountains which are presumably of Grenville age and sedimentary in origin (Engel and Engel, 1953, p. 1037).

The Mount Holly complex is often distinguished petrographically from younger rocks by its unique metamorphic characteristics. Mount Holly rocks preserve evidence of an earlier more intense metamorphism than that which produced chlorite and biotite in Paleozoic rocks. Grains of pyroxenes, intermediate plagioclase, and actinolitic hornblende are now nearly completely altered to appropriate minerals of lower grade. This alteration has left a characteristic stamp on Mount Holly rocks which, in the absence of structural discordance, commonly separates Mount Holly from younger rocks.

In spite of important differences in thermal history, however, the precise location of the unconformity is frequently very difficult. In addition to structural reasons to be discussed, differentiation of lowermost Cambrian from Precambrian rocks may be hindered by gross mineralogic similarity. Basal Cambrian is commonly composed of clasts derived locally. As both source and derivative rocks have been sheared and recrystallized during Paleozoic metamorphism, the final products may be nearly identical. Many mineralogic and petrographic criteria have been developed to meet this critical problem in the Green Mountain anticline (see for example, Thompson, 1950).

The contact of Mount Holly with Paleozoic rocks, though often obscure in detail, emerges with regional mapping as a surface of profound angular unconformity. On both east and west sides of the anticline Paleozoic beds trend nearly north; the dip is moderate east on the east limb and steep to overturned west on the west limb. In the center of the core of the anticline, however, the older rocks strike northwest and dip gently northeast (fig. 2), at least in the northern and central part of the region. Banding in the Grenville of the Adirondack mountains to the west has an easterly strike (Tectonic Map of the United States, 1944).

The Green Mountain anticline in the vicinity of Rutland plunges gently northward. Both it and subsidiary folds are overturned to the west and minor structures in Paleozoic beds are customarily in agreement with major folds.

Bouger gravity anomalies offer valuable indirect information about the Green Mountain anticline at depth (Bean, 1953; Diment, 1956). A strong relative gravity high at the axis of the anticline (fig. 1) is matched by a relative gravity low along the Middlebury syncline to the west. The interpretation is that deep crustal high-density layers are involved in the folding evident at the surface.

In the following discussion, "mantle" refers to the eastern and western

Fig. 2. The Rutland area, showing the Precambrian core of the Green Mountain anticline and mantling Paleozoic beds. The location of the area is given in figure 1, geologic structure sections BB and CC in figure 4, and details of the Saltash unit in figure 6. Note the northwestward trend of Precambrian marble and quartzite in the central part of the area, and the influence that this has had on the younger rocks of the Wilcox and Deers Leap synclines. Although locally somewhat variable, folds in Paleozoic rocks plunge gently northward in the Rutland area; note the contrast with folds in the Saltash unit.

sequences of Paleozoic rocks. The Precambrian Mount Holly is termed "core" or "basement (complex)".

INTERACTION OF BASEMENT AND MANTLE DURING FOLDING

Interaction of Paleozoic and Precambrian beds as their surface of separation is bent into an overturned anticline can be analyzed in two ways: (a) the imprint of Paleozoic deformation on the discordant Precambrian series, and (b) the influence of inhomogeneity of the basement in guiding or refracting Paleozoic movements.

READJUSTMENT OF PRECAMBRIAN ROCKS DURING YOUNGER FOLDING

There are numerous ways that the basement might readjust its shape to conform to geometrical requirements of flexural folding of the mantle, particularly during a deformation along directions sharply discordant with basement trends:

(a) The change from a flat to a strongly arched surface might cause wholesale fracturing and widespread granulation of the underlying core rocks, or,

(b) If conditions were such that material could flow plastically the basement might readjust its shape by some sort of irregular flow.

(c) It is possible too that the bulk of the basement might remain structurally more or less intact with all adjustment accommodated by intense fracture or flow near the surface of unconformity.

The situation in central Vermont is fortunately favorable to determine just how core and mantle did interact during very intense folding. Mapping

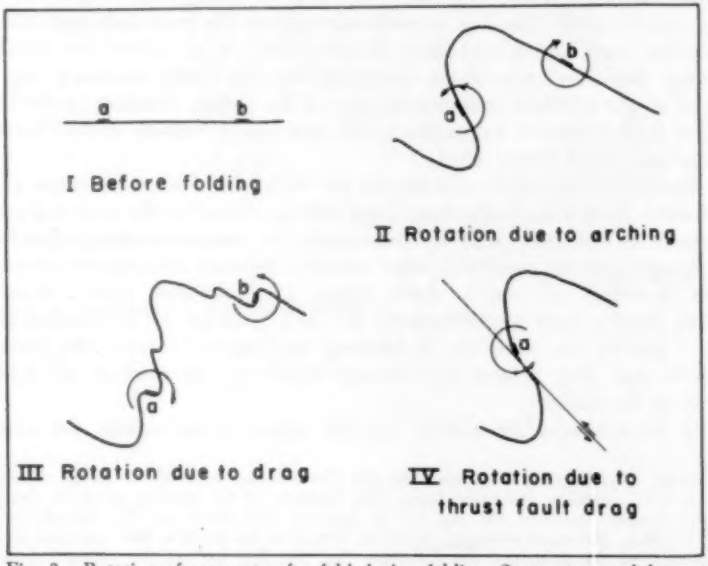

Fig. 3. Rotation of segments of a fold during folding. Segments *a* and *b* are traced through the various rotations usually required by flexural folding. Arrows show rotation of segments with respect to an initial horizontal surface. In natural folding these various rotations may be superimposed.

of key marble and quartzite beds disclosed in a rough way the structure of the core. Well away from the contact of basement and mantle there is remarkable uniformity of strike of the basement (fig. 2). As the unconformity is approached from below, however, basement rocks are intensely granulated and sheared, and the strike of key beds changes rather abruptly. Few faults were detected and these of small displacement. Intense deformation is restricted to a zone several thousand feet thick (measured stratigraphically) beneath the contact. This zone is more closely examined below.

There are certain aspects of flexural folding that are perhaps not immediately apparent. A segment of a limb of a fold rotates under the influence of several couples. In figure 3 such a fold is shown; the surface sketched might be considered the unconformity between basement and flanking strata. Segments on each limb of the fold rotate relative to an initial horizontal surface; first, due simply to the change in geometry of the surface; second, due to the drag of crestward moving overlying layers (expressed as minor folds); and third, due commonly to the effect of breaking and thrusting of the overturned limb. From figure 3 the directions of these various rotations are clearly not alike, which explains for example the inconsistencies often observed in minor fold geometry near break thrusts in sharply overturned limbs of folds.

It was not at first obvious which of these rotations would predominate in material just below the unconformity. Fortunately mapping provided the answer. For several thousand feet stratigraphically beneath the unconformity, Precambrian structures have been bent nearly conformable with younger rocks. The direction of bending shows the movement undergone by older strata in this part of the anticline; this is sketched in figure 4 for cross sections made at two locations across parts of the anticline. The relations shown are probably typical for this part of the Green Mountain anticline.

Rotation of Precambrian structures is quite spectacular on the west limb of the anticline; the material outlining the rotation is the Wilcox formation, which occupies a syncline within garnetiferous augen gneiss and amphibolite (fig. 4b). The west limb of this ancient syncline complete with associated minor structures has been rotated through about 90°. This can be recognized as the second rotation of figure 3: counter-clockwise looking north. The peculiar map pattern of the Wilcox formation reflects preservation of original drag folds.

In addition to this bending of the basement on the west limb, other movements are discernable. Rotation III of figure 3 appears in fact as a fault, with west side up relative to east side, cutting the west limb of the schist-dolomite syncline. Movement sense here is the same as that producing drag folds in overlying Cambrian beds (fig. 4b). Bending and faulting of the syncline were probably contemporaneous, as the two different deforming movements exist at the same time.

The basement of the east limb of the Green Mountain anticline (fig. 4a) has been bent in a counterclockwise direction looking north. This is recognizable as rotation III of figure 3. Bending is marked for a distance of several miles (fig. 2) by beds of quartzite, gneiss and marble.

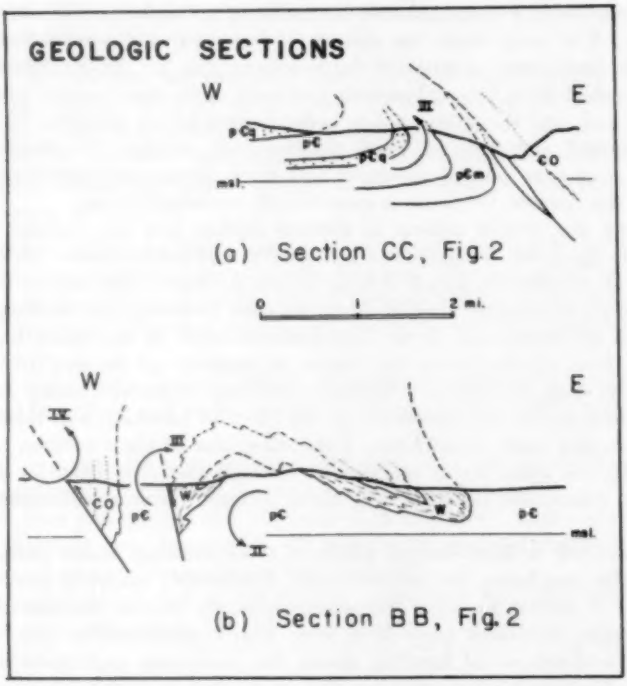

Fig. 4. Geologic structure sections BB and CC, referred to map, figure 2. Section CC is on the east limb and section BB on the west limb of the Green Mountain anticline. Arrows with Roman numerals, II to IV, identify rotations given in figure 3. p€q is basement quartzite, p€m is marble, and W the Wilcox formation; other symbols are as in figures 1, 2. Note particularly the extreme distortion of the west limb of the Wilcox syncline: minor structures and all have been bent nearly ninety degrees!

One of the great difficulties in mapping rocks near this folded unconformity is the almost constant parallelism of compositional banding which in Paleozoic and indeed in relatively undeformed Precambrian rocks is doubtless equivalent to bedding. However the true significance of composition banding in the highly disturbed marginal zone of the basement is not always clear. Banding may be primary or produced by shearing. If in part primary, as certainly shown by interbedded quartzite and marble, then a conclusion is possible regarding behavior of basement rocks here during folding of an unconformity: older layers are bent nearly parallel with the surface of contact. Both limbs of the Green Mountain anticline appear to show such behavior.

Comparison with other Areas.—As Thompson first noted[5], the major tectonic units (and their metamorphism) of the Western Alps are remarkably similar to those of western New England. The Aar, Mt. Blanc and Gotthard

[5] This similarity was first mentioned to me by Thompson following his early work in central Vermont.

massifs are very like the crystalline core of the Green Mountains; parautochthonous mantling strata of the Helvetides resemble at least geometrically Cambro-Ordovician beds of western Vermont. In his exhaustive work on the Alps, Heim briefly describes interaction of basement and mantling rocks (1921, Part II, p. 156). He observed that the schistosity of basement rocks of the north end of the Aar massif is bent and dragged nearly parallel to the contact with mantling beds. But even if dragging is absent, the normal foliation of the massif is obscured near the margin of the massif by a variety of new shear surfaces.

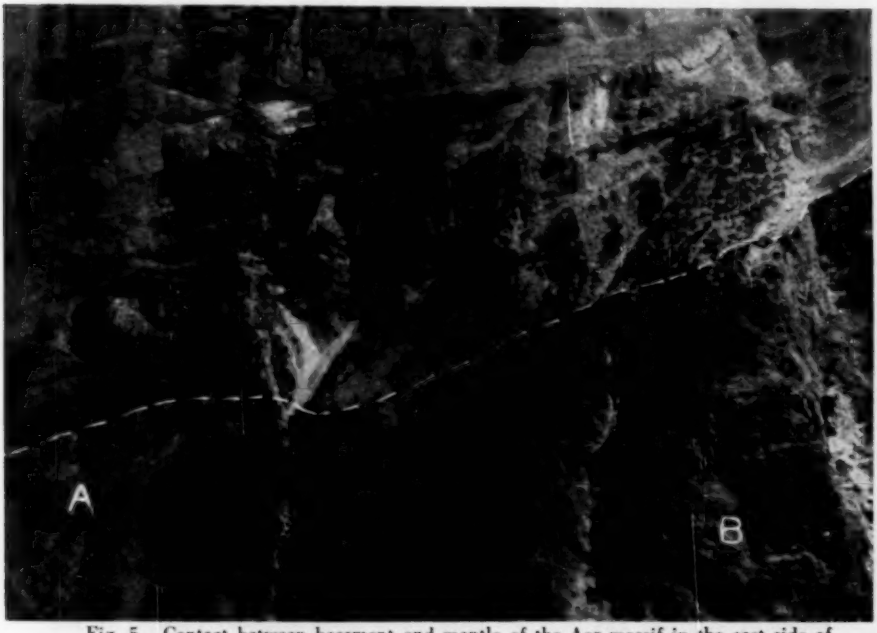

Fig. 5. Contact between basement and mantle of the Aar massif in the east side of Lauterbrunnental. White dotted line gives location of unconformity. Area has the following coordinates on the Jungfrau Sheet, 1:50,000, #264, Landeskarte der Schweiz: 154.2, 636.3. At B the normal vertical foliation of pre-Triassic gneiss is visible; at A this foliation is cut by shears parallel to the unconformity. Simplicity of structure in this photograph is misleading; enormous recumbent folds are nearby. The large drag fold above A gives some hint of this folding. The tallest trees, which are 50 feet high, give the scale.

At the north end of the Aar massif, Lauterbrunnental cuts across and clearly exposes the unconformity between Mesozoic dolomite, sandstone and phyllite, and underlying pre-Triassic crystalline rocks (fig. 5). The Mesozoic rocks dip gently north and rest on the crystalline rocks in which foliation is typically vertical (at B, fig. 5). Of particular interest here is the development of a foliation in the older rocks parallel to the unconformity as the unconformity is approached (at A, fig. 5). The rock breaks along surfaces spaced one-half to 2 inches along which there is slight secondary growth of mica and

commonly a slight shearing displacement. This foliation is presumably formed by shearing as it parallels bedding planes in the Mesozoic rocks that have acted as fault surfaces (bedding thrusts). Such foliation is commonly the predominant structural element, as for example in otherwise massive rocks, or where later recrystallization has accentuated foliation. Although this Alpine example is not bending of strata into conformity with the contact of the mantle, structural conformity with the mantle is achieved: foliation, the dominant planar feature of the older rocks is, in the vicinity of the unconformity, parallel to the unconformity.

Balk's classic study (1936) of Dutchess County, New York, is still another example of interaction of basement and younger rocks not unlike those already described. Thrust faulting has brought basement gneiss and younger schist into contact (Balk, 1936, p. 739, 741, fig. 30); foliation of the gneiss has been bent parallel with the fault as the fault is approached. Now faulting is not obviously important in the Rutland area but bedding planes in the younger rocks have consistently acted as surfaces of slip, as shown by studies of mineral and grain orientation (Brace, 1954). Therefore the effect on the older rocks should be the same, either for a single fault, or for many small shears on bedding. The result is again parallelism of planar fabric in older and younger rocks.

Recent work of Hudson (1955) summarizes certain aspects of this interaction problem for rocks deformed in a much different environment at various localities in California and the Rocky Mountains. He studied the folding of basement crystalline rocks in what was termed "the zone of fracture" where mantle sediments were known to be thin and unmetamorphosed. Hudson's work is particularly interesting for the light shed on the actual mode of readjustment of basement to the folding. Again readjustment is a marginal feature (Hudson, 1955, p. 2045-2046). Folding of basement-mantle contact, especially of granitic basement, is tolerated by a thin *zone* of crushed material just below the unconformity. It is presumed that the thickness of the crushed zone varies inversely as the radius of curvature of the fold.

Summary.—Formation of a large anticline about a core of crystalline metamorphic rocks has bent the foliation or compositional banding of the core rocks nearly parallel with younger mantling strata. Bending, dragging or shearing is limited to a marginal zone which leaves the interior of the core relatively intact. Concordance of structure is produced by shearing as well as bending; shears typically form in older rocks parallel to the contact. Adjustment of basement to shallow folding is also limited to a marginal zone.

EFFECT OF BASEMENT INHOMOGENEITY ON FORMATION OF MANTLE STRUCTURES

One usually expects the formation of younger structures to be strongly influenced by inhomogeneity of the rocks on which they lie and against which they are unconformable. In too few situations can this influence be specified, for basement rocks are usually inaccessible. In the Green Mountain anticline, basement influence is manifest in both (a) the trends of several younger folds, and (b) the development of both depositional and deformational pattern of a unique lens of late Precambrian rock.

Trends of Younger Folds.—The importance of Precambrian structural trends in guiding the geometrical development of the Green Mountain anticline is probably small; the anticline trends north whereas basement structure both in Vermont and New York trends northwest to west. However the details of this geometry may owe much to control by the basement.

Two large synclines traceable within the central part of the core are distinctly parallel with basement trends. These are the Wilcox syncline previously discussed, and the Deers Leap syncline in the northern part of the map area (fig. 2). Whether this parallelism is caused by depositional or entirely deformational control is not known, but it is clear that the movements which are responsible for this geometry must have been greatly affected by basement structure. Although the Precambrian had already been strongly recrystallized, great physical anisotropy would have still existed, as for example in beds of quartzite within gneiss. Surfaces of compositional banding probably would have acted as slip surfaces in the zone of most intense shearing near the contact with younger sequences. This influence is particularly apparent where massive Precambrian quartzite and quartz-schist occur near the unconformity.[6] The movements responsible for forming the younger folds, although of Paleozoic age and generally oblique to basement structure, have apparently been resolved near the unconformity into directions within the basement banding.

Depositional and Deformation Control of the Saltash Unit.—The Saltash formation occurs between older gneiss and quartzite, and Paleozoic phyllite and volcanic rocks (fig. 6). The Saltash is here included in the "mantle", for it is geometrically like the Paleozoic if not stratigraphically of Paleozoic age. In the underlying basement there is a significant arrangement of rock units. As indicated in figure 6 large masses of quartzite occur just to the north and south[7] of the late Precambrian wedge; between these buttress-like masses schist, augen gneiss and marble crop out. The present characteristics of the Saltash may be due to (a) basement control of deposition, and (b) the effect of basement inhomogeneities on structural evolution.

Depositional control is not clear, owing primarily to lack of information beyond the area studied here in detail (Brace, 1953, p. 51-54). However, the Saltash probably owes its lens-shaped form to the topography of the basin in which it was deposited. This topography consisted, at the level now exposed by erosion, of hills underlain by quartzite, and a trough more deeply eroded in softer schists and marbles. Sediment filled the trough and covered the quartzite hills; lithification followed and the rocks were then uplifted. Erosion cut downward and removed the Saltash from all but the deeper parts of the original basin: the trough formed in gneiss and marble. Accordingly only a lens of the Saltash formation remains.

[6] Occurrence of massive Precambrian quartzite near the unconformity usually signals disorder in younger overlying beds. The regional offset of the Green Mountain west limb just north of Rutland (fig. 1) is attributed to the disturbing effect of abundant basement quartzite at this latitude (Brace, 1953, p. 71).

[7] The southern quartzite mass is described by Thompson (1950).

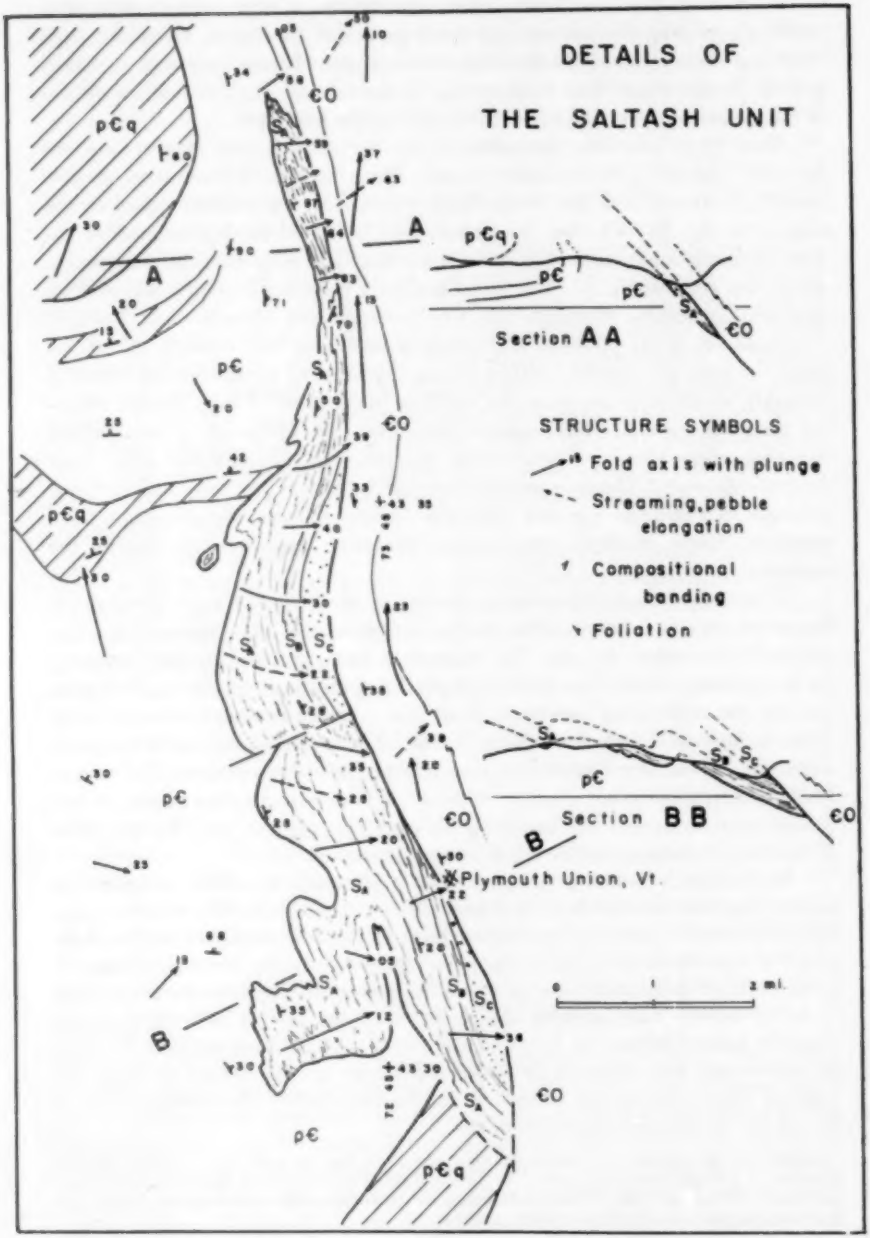
DETAILS OF
THE SALTASH UNIT
Section A A
Section B B
STRUCTURE SYMBOLS
Fold axis with plunge
Streaming, pebble elongation
Compositional banding
Foliation
Plymouth Union, Vt.
pCq
pC
€O
A
B
0
1
2 mi.

Fig. 6

The pattern of minor structures in the Saltash formation is even more unusual than the shape. Two systems of structural elements are present, one a schistosity and folding parallel to those elements in overlying Paleozoic beds, and an older and more obscure system of folding and lineation. The first system clearly stems from the major folding responsible for the Green Mountain anticline. However, the second system suggests a much different history. The older folds plunge gently eastward in the direction of dip of the bedding and are parallel to a strong mineral lineation and to the direction of enlongation of stretched cobbles. Plunges of folds diverge slightly in an easterly direction. Furthermore, the folds are asymmetric; the axial planes approach the attitude of bedding.

Steeply plunging folds with direction of plunge subperpendicular to the regional trend of beds could have been formed in several ways.[8] One explanation is that the Saltash is the west limb of a large isoclinal fold which plunges eastward as do the minor folds which we observe. However, this scheme introduces more complication than clarification. Generation of such a structure would require either a horizontal couple or rotation of the major fold after formation by a vertical couple. Either possibility is unlikely; such movements are inconsistent with the regional movement picture. In addition, no imprint of such bizarre movements appears in the basement immediately beneath the Saltash. Finally, the pattern of the minor folds cannot be integrated into a consistent pattern of shear as it would if this were part of a large isoclinal fold.

The unique linear structure of the Saltash reflects in a striking way inhomogeneity of the basement rocks. Normal movement during formation of the Green Mountain anticline was an east over west shearing at this part of the anticline; small folds are arranged parallel to the large fold and reflect this normal movement. Now the additional factor here is that resistant masses in the basement interfered with a uniform movement of layered units westward up the eastern limb of the anticline. Converging movement caused a limited shortening parallel to the fold axis. The quartzite masses north and south of the Saltash (fig. 6) have not moved toward one another but have in effect produced an north-south shortening by acting as constrictions in the overall westward flow of plastic rock masses. In response to this subsidiary north-south shortening, numerous small folds with axes trending eastward developed in the Saltash.

Folding about axes parallel to the direction of maximum extension suggests similarity of the Saltash evolution to that of cylindrical plutons (Balk, 1937, p. 86) and salt domes (Balk, 1949). These likely analogues have been

[8] These are *a*-folds in petrofabric terminology; see Fairbairn (1949, p. 222) for review of the literature.

Fig. 6. Details of the Saltash unit. The three-part Saltash is shown in relation to Precambrian basement quartzite (p€q) and schist, marble and gneiss (p€), and argillaceous and volcanic Cambro-Ordovician rocks. The southern quartzite mass (p€q) is only approximately located, based on Thompson (1950). Further details of the stratigraphy and metamorphism are given in Brace (1953). S_A is grit, S_B black phyllite and dolomite and Sc vitreous white quartzite. Note the predominant fold trend in the Saltash, approximately normal to folds in overlying Cambro-Ordovician.

constricted in two directions normal to the direction of maximum movement and as a consequence have formed folds parallel to the direction of maximum movement.

The origin of the linear structure of the Saltash is also perhaps similar to that of structures common at thrust faults[9], such as folds, rodding and mineral lineation. The small folds of the Saltash however are commonly overturned and even isoclinal and therefore correspondence is somewhat weakened; furthermore, it is often in doubt that shortening normal to the direction of thrusting has been the cause of the folds at thrusts. This mode of origin is strongly suggested for the Saltash.

Basement Control in Western Norway.—Basement influence during deformation of overlying rocks has often been advocated to explain details of rock geometry. Particularly detailed work has been done by Kvale in the Bergsdalen area of Norway (1948), also with folded and overthrust metamorphic rocks. Pronounced differences in competence exist there between such rocks as diorite plutons and schist. From careful study of major and minor structures Kvale could say:

"A solid body of rock, especially a pluton, will tend to influence the direction of the linear structures in the less solid surrounding rocks in such a way the lineation or axes of folds or both these structures become more or less parallel to the boundary of the solid body (Kvale, 1948, p. 248, fig. 6)."

Now the Saltash linear structure has just this relation to the basement quartzite buttresses (fig. 6). Further, the divergence of direction of plunge of the folds is in striking agreement with the geometry of folds in the Bergsdalen area. As the quartzite buttresses are neared, folds become more nearly parallel to the contact with the buttress.

Summary.—Inhomogeneity or anisotropy in rocks beneath an unconformity can strongly influence structural trends formed in younger beds. Folds and certain other lineation may leave the normal orientation and approach more closely the direction of contact with relatively competent or resistant beds in the basement. As in the Rutland area, this departure from normal trend may sometimes produce folds perpendicular to the major anticlines and synclines.

CONCLUSIONS

In part of the Green Mountains of central Vermont, structural evolution of Paleozoic rocks and a structural modification of part of the Precambrian basement were closely related. Both occurred during folding about an axis oblique to Precambrian trends. During this (Taconic) folding, the basement adjusted to the imposed change of shape in a marginal zone, several thousand feet thick, near the unconformity. In this zone Precambrian beds were either dragged nearly parallel with younger strata, or sheared along surfaces parallel to the unconformity.

Inhomogeneity in basement rocks produced certain structural peculiarities in mantling beds. Large masses of quartzite acted as buttresses and caused

[9] See for example, Fairbairn, 1949, p. 219-222.

folding along axes which are nearly parallel to their margins; these folds are particularly striking when sub-perpendicular to the normal fold direction.

The relative importance of either (a) bending of basement or (b) deflection of younger beds from normal Taconic trend depends of course on comparative competence or resistance. Precambrian schists and marbles are bent and dragged, whereas massive quartzites act as buttresses. On the other hand the role of the Paleozoic rocks is less important. Near the unconformity variations in thickness and lithology of Paleozoic rocks have little effect. For flexural folding and dragging to have occurred at all, however, it was of course necessary that the Paleozoic rocks possessed some strength; these rocks must have already been fairly coherent at the onset of the principal Taconic deformation. Nevertheless, the ruling factor here was certainly variations in strength of the basement rocks.

EXTENSION TO REGIONAL PROBLEMS

Mapping of basement complex in a region such as the Green Mountains has value in that it helps to explain certain details of the geometry of younger rocks. Also, light may be thrown on more far-reaching tectonic problems. It may be possible by study of the basement complex to place a limiting figure on overall shortening of a folded system of rocks. In this study, with readjustment of the basement restricted to relatively thin zones marginal to the unconformity, we are at least limited to comparatively slight lateral shortening of the core rocks. Otherwise, there is absolutely no limit to the distance which could have originally separated the east from the west limb of the present fold.

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IDIOMORPHIC ZIRCON, KEY TO VOLCANISM IN THE LOWER TERTIARY SANDS OF CENTRAL TEXAS

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ABSTRACT. Lower Tertiary sands of central Texas consist of a section of sands with very little volcanic material, succeeded by a section of richly volcanic sediments. Zircon types have been counted from ten of these sands and found to occupy characteristic fields when plotted on a triangular percentage diagram utilizing the three poles (1) idiomorphic plus hypidiomorphic, (2) angular to subangular, and (3) round to subround. The chief difference between immediately adjacent formations is due to the change in proportion of rounded to angular zircons, which appears to be largely a function of source area modified only slightly by abrasion in the final environment of deposition. The richly volcanic beds have a significantly higher proportion of idiomorphic zircon, and even minor spurts of volcanism can be recognized by abnormally high idiomorphism values. Zircon idiomorphism is nearly independent of zircon length, except for samples having zircons averaging smaller than 0.08 mm which show too low idiomorphism.

INTRODUCTION

Interest in the geologic significance of zircon has been heightened greatly by the researches of Poldervaart (1955, 1956), who has dealt extensively with length and width of zircon grains. In this study, we have investigated another aspect of zircon shape, specifically idiomorphism.

For several years heavy mineral suites from Tertiary sediments of the Texas Gulf coast have been studied as a part of graduate thesis work and petrographic reconnaisance at the University of Texas. These suites consist largely of kyanite, staurolite, zircon, tourmaline, garnet, rutile, and opaque minerals. One of their striking features is the great diversity of zircon types. Any single mount may contain perfectly idiomorphic and complexly-faceted grains, nearly spherical well-rounded grains, elongated narrow rods, or broken angular fragments (fig. 1). Many grains contain abundant inclusions and a small proportion of pink to deep purple zircons occurs. This diversity is in accordance with the extremely complex provenance of Texas Tertiary sediments; for example, sands of the Lower Eocene (Todd, 1956; Todd and Folk, 1957; Adams, 1957; Harris, 1957) were derived from terranes containing rocks of low- to high-rank metamorphic, plutonic igneous, silicic and mafic volcanic, hydrothermal and pegmatitic, and older sedimentary origin. The source area for these sediments apparently included a broad section of the continent, ranging from the southern Appalachians through the mid-continent region to south Texas.

The first clue to the possible significance of the idiomorphic zircons present in these mounts was discovered when a heavy mineral sample of the Catahoula sand (Oligocene ?) was examined during routine reconnaisance and found to contain a strikingly high proportion of beautifully idiomorphic zircon. The light minerals of this sand consisted almost entirely of fine-sand size, slightly abraded volcanic glass fragments. Hence the supposition arose that the abnormally high proportion of idiomorphic zircon was likewise the result of volcanism. If this were so in the highly volcanic Catahoula, then were the scattered idiomorphic zircons in the non-volcanic part of the section the result of more sporadic volcanic activity?

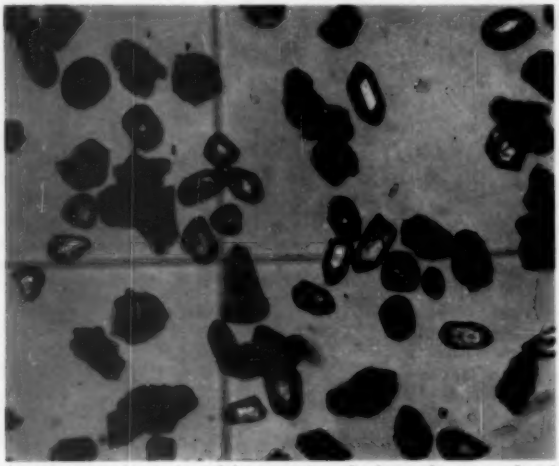

Fig. 1. A representative suite of heavy minerals from the non-volcanic part of the section (this is the Queen City sand). Most of the black grains are opaques; note the diversity of zircon types, ranging from perfect idiomorphs to completely rounded grains. Magnified ×95.

To check this hypothesis, we determined to study zircon types throughout the lower Tertiary section of Bastrop and Fayette Counties, southeast of Austin in central Texas. This area contains excellent exposures of Wilcox (Eocene) through Catahoula (Oligocene ?) beds. An abbreviated section, showing only the sands used in this study, is shown in figure 6. Petrographically, this section can be divided into two major parts: a lower, relatively non-volcanic part including Wilcox through Sparta sands; and an upper, increasingly volcanic part including Yegua through Catahoula strata.

Sands in the lower part of the section are rich in metamorphic rock fragments (slate, phyllite, schist, metaquartzite) and average in the subgraywacke category, although some range into orthoquartzite, subarkose or feldspathic subgraywacke fields (Folk, 1954) in response to shifting source areas, changing climate or amount of abrasion. Volcanism did occur sporadically in this time span, although most sands contain well under 2 percent of recognizable volcanic materials. Scattered analyses show that the intercalated shales are normally not bentonitic and consist chiefly of illite, kaolinite, or mixed-layer montmorillonite.

Volcanism started to be a significant contributor to sediments during deposition of the Cook Mountain formation (lying between the Sparta and Yegua sands), which contains some thick bentonitic clays. Because sands are uncommon in the Cook Mountain formation of Bastrop County, no counts of Cook Mountain zircons were made. Further up the section volcanism increased steadily to culminate during deposition of the Manning and Catahoula formations, which consist almost entirely of bentonite, vitric tuff, and volcanic sands.

The plan of attack for this study was to correlate idiomorphism of zircon with relative volcanicity of the sands as estimated from independent data. The following criteria have been used as indices of volcanism:

1. Abundance of volcanic glass shards and abraded glass grains, or recognizable sand-size fragments of holocrystalline aphanitic volcanic rock (identified by presence of felted plagioclase laths).

2. Presence of plagioclase in amounts greater than about one-tenth of the total feldspar percentage. Non-volcanic Texas Tertiary sands contain almost no plagioclase (under 0.1 percent) although they generally run 5 to 10 percent potash feldspar. In this area the presence of more than a trace of plagioclase has proved a very useful volcanic indicator, especially if the crystals are zoned.

3. Idiomorphic quartz phenocrysts, consisting of doubly-terminated hexagonal bipyramids of high-temperature quartz, lacking prism faces and usually showing some rounding of corners and embayments because of resorption in the magma. These crystals are easily recognized under the binocular microscope or even in the field by their shape, extreme polish (produced by resorption), brilliant luster and extreme clarity. They are inclusion-free except for occasional negative crystals also showing hexagonal bipyramid form.

4. Sub-hexagonal biotite flakes. In the normal, non-volcanic Tertiary sands mica is abundant (1-2 percent) but there is 50 to 100 times as much muscovite as biotite; in the volcanic beds the ratio approaches 1:1 and in some of these thin beds biotite is virtually the only mica present.

5. Apatite in the heavy mineral separates. Normal Tertiary sands here contain no apatite; in many of the volcanic beds, apatite occurs as large, slightly rounded prisms. Similar grains occur in the volcanic province of the Cretaceous Woodbine sand of northeast Texas (Cotera, 1956).

6. Presence of bentonites or bentonitic clays interbedded with the sands.

The following sands were examined in this study. Descriptions follow the system of Folk (1954, 1956) in which sediments are described on a two-fold basis of textural maturity and mineral composition.[1]

[1] The four stages of textural maturity are (1) immature, the sand contains over 5 percent clay and the grains are usually poorly sorted and angular; (2) submature, the sand has been winnowed of clay but the grains are still not well sorted (σ over 0.50ϕ) and are angular; (3) mature, the sands are well sorted but the grains are still angular to subangular; and (4) supermature, the grains are well sorted and subround to round, with Wadell roundness over 0.35.

The mineralogic clan is determined on percentage composition of the silt-sand-gravel fraction, ignoring clay matrix. Graywacke is a rock containing over 25 percent micas and metamorphic rock fragments (including stretched metaquartzite) and less than 10 percent feldspars; subgraywacke contains 5 to 25 percent metamorphic constituents and less than 10 percent feldspar. Arkose contains over 25 percent of igneous material including feldspar (K-feldspar or plagioclase) plus all igneous rock fragments (silicic or mafic, extrusive or intrusive), and less than 10 percent metamorphic material; subarkose contains 5 to 25 percent feldspathic and igneous material, and less than 10 percent metamorphics. Rocks consisting of mixed metamorphic and igneous material are impure arkose, feldspathic graywacke, and feldspathic subgraywacke. An orthoquartzite contains less than 5 percent feldspathic and less than 5 percent metamorphic material, hence is over 90 to 95 percent quartz and/or chert. Almost any composition may be present in almost any stage of textural maturity (Folk, 1956) and the definitions above apply regardless of color, induration, geologic age, sedimentary structures, or grain size.

Simsboro sand, Wilcox Group. Adams (1957) shows that the Simsboro is a very friable, gray-white, well-sorted fine sand. It is a texturally mature orthoquartzite to quartzose subgraywacke, deposited as a transgressive beach sand. Abundant detrital chert show that it was derived very largely from older sedimentary-rock sources. Only slight volcanic activity took place, the only evidence being a few hexagonal biotite flakes, a trace of apatite and very rare quartz phenocrysts.

Sabinetown sand, Wilcox Group. Harris (1957) determined that the average Sabinetown sample is a light brown to gray, very friable, moderately sorted silty very fine sand. It is a submature to mature subgraywacke, in places heavily cemented with calcite. The Sabinetown is a product of rapid deposition in a very nearshore marine deltaic-lagoonal-neritic complex, and contains abundant lignite and glauconite. The high content of slate and phyllite fragments indicates derivation largely from a low- to high-rank metamorphic terrane with some older sedimentary rocks. A fair amount of volcanism occurred: apatite and hexagonal biotite are common in many of the heavy mineral slides, and most thin sections contain plagioclase and basic volcanic rock fragments in amounts of 1-2 percent.

Carrizo-Newby sands, Claiborne Group. Todd (1956; also Todd and Folk, in press) has found that these sands are normally yellow to white, friable units that range from moderately sorted granular medium sand (lower Carrizo) to moderately- to well-sorted fine sand. Occasional zones, especially the lower Newby, are very hard and deep red because of thorough cementation by hematite. They are classified as submature to mature subgraywacke to orthoquartzite; the source area was largely low- to high-rank metamorphic with some older sediments. The Carrizo is chiefly a river-channel sand while the Newby is nearshore marine and contains considerable glauconite and authigenic feldspar toward the top. Volcanism was weak, the only evidence being considerable amounts (up to 4 percent) of quartz phenocrysts especially in the basal Carrizo, but there is no biotite or apatite and virtually no plagioclase or volcanic rock fragments.

Queen City sand, Claiborne Group. Work in progress by Callender shows that the Queen City is a friable, pale yellow, well-sorted very fine sand. It was deposited under fluctuating nearshore marine, deltaic, lagoonal and river channel environments, and is classified as a mature subgraywacke to orthoquartzite. The source area consisted of low- to high-rank metamorphic rocks plus older sediments. Volcanic activity throughout the bulk of the formation was virtually nil, but there are a few scattered quartz phenocrysts and one six-inch bed rich in hexagonal biotite.

Sparta sand, Claiborne Group. No detailed study has yet been made; reconnaisance indicates it is a friable, white well-sorted very fine sand, probably a mature orthoquartzite. No evidence of volcanism has been noticed.

Yegua sand, Claiborne Group. Examination of scattered samples indicates that it is a well to moderately sorted fine sand, probably a submature to mature plagioclase subarkose. Zoned plagioclase and volcanic rock fragments

are common in thin sections, and there is some hexagonal biotite; clays are richly bentonitic in places.

Jackson Group sands have not been studied in detail; the Caddell and Carlos are fine to very fine sands containing considerable plagioclase and some volcanic rock fragments and quartz phenocrysts. Great thicknesses of bentonite, of considerable commercial value, occur in the Caddell formation. The Manning formation is almost entirely vitric tuff, bentonite, and ash with few sands.

Catahoula formation, Oligocene (?). The Catahoula has not been studied precisely, but it is almost entirely volcanic with abundant bentonite and tuff; many beds of sand consist of nothing but abraded grains of volcanic glass, others contain abundant zoned plagioclase, sanidine and volcanic rock fragments.

Summary. The section from Simsboro through Sparta is dominantly non-volcanic, except for a very minor spurt of intermediate to basic volcanism during the Sabinetown; extensive volcanism began during Cook Mountain or Yegua deposition and continued in ever-increasing amounts through the Manning and Catahoula, so that the latter two formations are almost entirely thick accumulations of volcanic debris.

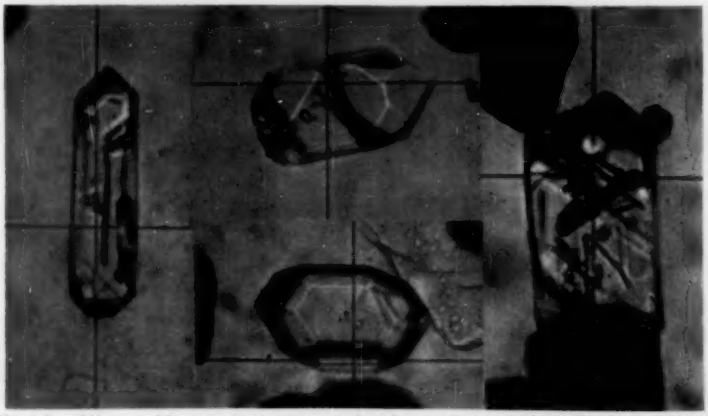

Fig. 2. Idiomorphic zircon grains, with planar faces and perfectly straight and sharp edges. Two of the zircons contain abundant tubular fluid inclusions. Magnified ×450.

METHOD OF STUDY OF ZIRCON

Heavy minerals were separated with bromoform and separatory funnel as described in Krumbein and Pettijohn (1938, p. 343-344), and all heavy minerals (including zircon) were mounted together. Usually fifty zircon grains were counted per slide, by using a mechanical stage and making equi-spaced traverses covering the entire mount. Every zircon grain passing under a ruled grid in the ocular (grid 1.1 mm square) was tabulated. Since the major aim of this study was to study the significance of idiomorphism, the basic classification used was (1) idiomorphic, (2) hypidiomorphic, (3)

xenomorphic angular to subangular, (4) xenomorphic round to subround. These classes were operationally defined as follows:

1. Idiomorphic. Crystal faces perfectly smooth, mirror-like. Edges and corners of faces perfectly sharp and straight. No evidence of any abrasion whatsoever (fig. 2).

2. Hypidiomorphic. Crystal outline still plainly visible, but corners are slightly rounded and faces are scarred. Long prism faces still present and not much modified (fig. 3).

3. Subround to round (xenomorphic). Little or no evidence of crystal outline, corners rounded (fig. 4).

4. Subangular to angular (xenomorphic). Little or no evidence of crystal outline, corners angular (fig. 4).

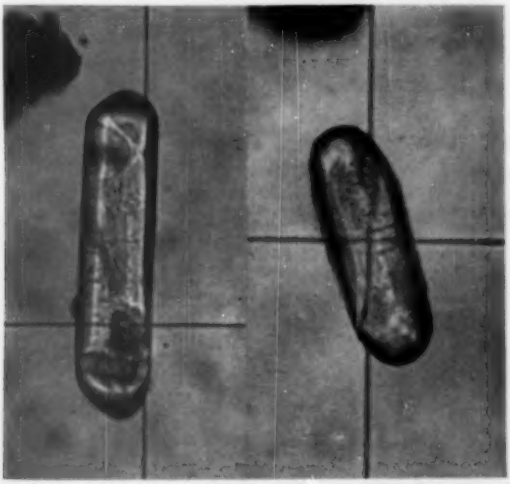

Fig. 3. Hypidiomorphic zircon grains. Prism faces are still pretty straight, although noticeable rounding has occurred; the smaller grain is almost at the boundary between "hypidiomorphic" and "round" classes. Magnification ×450.

Allotting grains to the above shape classes is an intensely personal operation. Poldervaart (personal communication) has pointed out that he would classify the grains of our figure 3 as "subrounded" because his work emphasizes crystal terminations, whereas we have classed these grains as "hypidiomorphic", stressing the long prism faces that are nearly unaltered. Consequently if other workers counted the same samples we worked on, they would undoubtedly obtain different percentages of zircon types; but the relative differences between formations should still hold up. For example, all workers should find that the Yegua formation has more idiomorphic zircons than the Sparta sand, although one man might get 60 percent versus 40 percent while another man got 40 percent versus 25 percent.

Since the estimations of degree of idiomorphism are so subjective, it would have been extremely easy for operator bias to creep in. Because all the counts were made by Callender, there is no between-operator variation. But

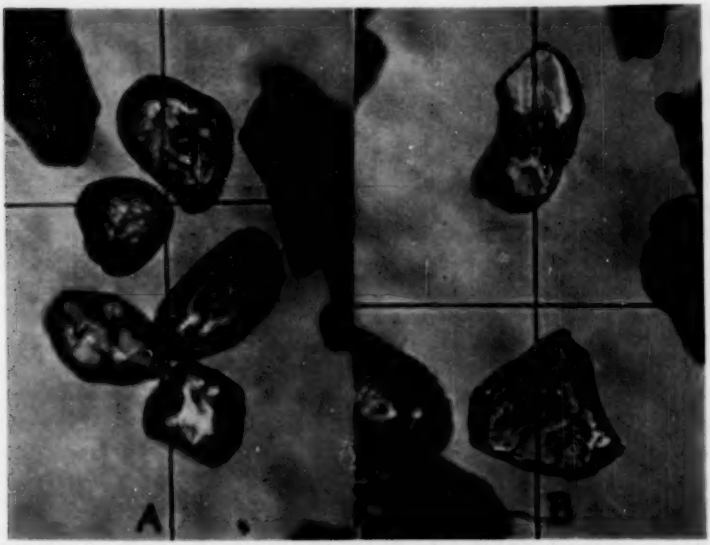

Fig. 4. Xenomorphic zircon grains. In photograph A, the top three grains are classed as "round", and the bottom two grains are on the borderline between "round" and "angular". In photograph B, the lower grain is "angular" and the upper grain is borderline between "angular" and "round". Magnified ×450.

if Callender had known at any time that the particular heavy mineral slide he was counting came from a formation in the volcanic part of the section, his count would doubtless have been influenced toward higher-than-true idiomorphism by the always-present subconscious tendency to "prove up" the hypothesis. Hence special precautions were taken to insure against this bias; the slides were given meaningless code numbers and completely intermixed so that slides from the same formation were done at different times. Thus at no time did Callender know what formation he was working with in the zircon counts, until the very end when all analyses were plotted up.

While primary importance was given to shape, zircons were also classified on the following bases: (1) elongation, with a separate tabulation being made of idiomorphic and hypidiomorphic zircons with length/width ratios greater than 2.5:1 (xenomorphic zircons were not so classified); (2) presence or absence of prominent inclusions (fig. 2); the inclusions were large, tabular, randomly oriented vacuoles or canals, apparently filled entirely with either gas or liquid (i.e., no bubbles were seen within the vacuoles); (3) color, with pink to purple varieties being separately tabulated.

Study of these other properties failed to reveal any obvious changes up section or between formations, although it is possible that a much more detailed study emphasizing these properties might reveal small differences. About 10-15 percent of the idiomorphic to hypidiomorphic zircons were more elongate than 2.5:1, this ratio remaining fairly constant up-section. About 70 percent of the idiomorphic to hypidiomorphic zircons contained prominent

fluid inclusions, but only 40 percent of the xenomorphic ones contained them. No stratigraphic differences were observed in inclusion content. Almost all zircon grains were colorless with only 2-3 percent being purple. The purplish zircons tend to be stubby (shorter than 1.5:1) and slightly rounded hypidiomorphic to subround, although a few grains are idiomorphic or angular xenomorphic. One gets the feeling that the purple zircons are much more uniform in shape than the other types, implying that they may have come from a single source which continued to shed them during deposition of all the formations studied. No significant stratigraphic change was noted. Occasional yellow and brown zircons were noted, but these formed no more than one percent of the total.

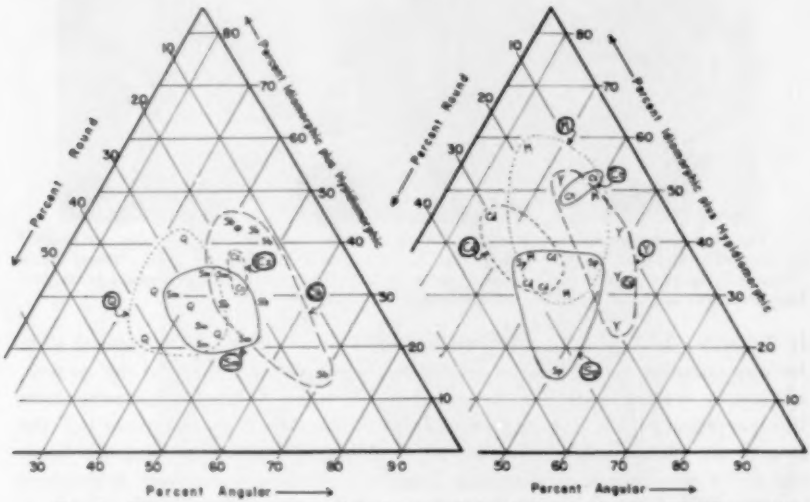

Fig. 5. Triangular percentage plot of proportions of zircons that are (1) Idiomorphic plus hypidiomorphic, (2) Round to subround xenomorphic, and (3) Angular to subangular xenomorphic. The composition of each 50-grain count is shown by an appropriate letter symbol (e.g. Sp for each Sparta sample, Q for each Queen City sample, etc.). All samples from a given formation are enclosed by a line to define the field of variation. Both triangles are the same; it was necessary to plot the older formations on one triangle and the younger ones on another to make the diagram legible and avoid excessive overlap. Formation symbols are as follows: left-hand triangle, Sm, Simsboro; Sb, Sabinetown; Cz, Carrizo-Newby; Q, Queen City (the asterisk at 43% I, 42% A represents a sample from the thin volcanic bed within the Queen City). Right-hand triangle, Sp, Sparta; Y, Yegua; Cd, Caddell; M, Manning; Cs, Carlos (one sample only); Ct, Catahoula. Observe the generally higher positions on the idiomorphism scale occupied by the highly volcanic formations (Yegua and younger). The difference in proportion of round versus angular zircon is striking in the case of the Yegua and Caddell, two immediately adjacent sands.

ZIRCON TYPES AS A CORRELATION TOOL

The zircon counts have been plotted on a triangular percentage diagram (fig. 5) using the three poles (1) idiomorphic plus hypidiomorphic, (2) xenomorphic angular to subangular, (3) xenomorphic round to subround. From each hand specimen, a count of approximately 50 zircons was made;

the composition of each count is represented by a single symbol on this triangle, and all samples from a given formation are enclosed by a line to show the field of variation encountered. Although there is some degree of overlap, in general there is a good differentiation of formations on this plot, and each formation occupies a rather closely condensed field. This is particularly encouraging because a new batch of samples was collected and counted almost a year after the first counts had been made, and the new counts fell right in the proper fields. It would appear from this diagram that in general a count of 50 zircons from each of five representative samples (a total of 250 zircons) would suffice to identify the formation. This would be particularly useful in places like Bastrop County, Texas, where sands that appear so similar on surface outcrop are often chopped up badly by faulting and it is sometimes difficult to know which sand one is in. For example, comparing the Sparta and Yegua zircons, the χ^2 test gives $P = .02$ (i.e., there is only a 2 percent chance that they have the same zircon population), with greatest differences due to the larger proportion of rounded zircons in the Sparta versus angular ones in the Yegua (see fig. 5). Comparing the Yegua and Caddell counts, $P = .004$ (only 4 chances in 1000 of their having the same zircon population); again this is due almost entirely to the greater angularity of Yegua zircons, as here both formations have essentially similar proportions of idiomorphic zircon (fig. 5). The latter two immediately adjacent formations are so different in zircon morphology that a count of 100 grains per formation should suffice to tell them apart with $P = .05$, assuming the differences we have found hold up. Sometimes this method fails, however; the Queen City and Sparta sands appear to have very similar zircon populations with $P = .30$.

Judging by the contributions to χ^2, differences between *immediately adjacent* formations are due largely to differences in proportions of angular vs. rounded zircons. The proportion of idiomorphic zircon, on the other hand, changes more slowly and regularly up section (save for occasional "spurts"), hence adjacent formations differ little in idiomorphism. There are two possible causes for the difference in proportions of angular vs. rounded zircons. First, the cause might be environmetal factors, e.g. the fact that some formations were laid down as beach sands wherein zircon was being actively rounded, while other formations are fluvial or neritic sands wherein no zircon rounding was being accomplished. Secondly, the difference might be due to inheritance, i.e., different proportions of angular vs. rounded zircons were contributed to the sediment in response to changes in location or composition of the source areas, and little or no modification of the zircon shapes occurred by abrasion in the final environment of deposition.

Evidence enabling a decision between these two possibilities is mixed. The following points favor the hypothesis that the environment of deposition was able to abrade zircons and hence increase the proportion of rounded grains: (1) the highest proportion of rounded grains is found in those sands that, according to our present data, were deposited under the most marine or littoral conditions: the Simsboro, Queen City and Caddell formations; the more non-marine (or low-energy deltaic) formations have more angular

zircons. (2) Adams (1957) shows that nearly all idiomorphic tourmaline prisms and quartz grains in the Simsboro show noticeable edge-rounding, while Harris (1957) demonstrates that tourmaline prisms have completely sharp edges and faces, and quartz grains are similarly sharp-edged in the deltaic Sabinetown formation. This strongly implies that the beach-littoral environment that prevailed during deposition of the Simsboro was capable of imparting slight but noticeable rounding to such hard minerals as tourmaline and quartz, and presumably zircon would also be affected.

The following points favor the hypothesis that the difference in zircon roundness is strictly inherited from changing source terranes: (1) very little rounding of quartz has been shown to have occurred in the Tertiary sands of Texas, with the exception of the Simsboro. Even such soft minerals as feldspar and kyanite generally show angular to subangular shapes, with only occasional edge-rounding. (2) The Simsboro sand, upon which the major case for environmetal rounding rests, is derived very largely from older sedimentary rocks and has, therefore, a very high proportion of *inherited* rounded zircons. Unfortunately the petrography of the Queen City and Caddell sands has not been thoroughly studied yet and the relative volume of reworked sedimentary material has not been determined.

With present sketchy evidence, the following tentative conclusions seem to be developing: the main control on the ratio of rounded to angular zircon grains is the changing zircon population contributed from differing source terranes, specifically the proportion of older sedimentary rocks exposed in the source. However, in certain formations deposited under unusually potent littoral conditions (e.g. Simsboro), the mean roundness of the inherited zircon

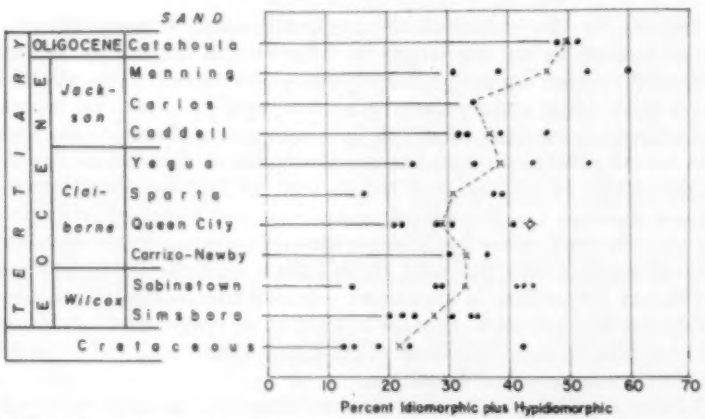

Fig. 6. Stratigraphic change in the proportion of idiomorphic and hypidiomorphic zircon. Each dot represents a 50-grain count; averages for each formation are denoted by crosses connected with a dashed line. The open circle represents a sample from the thin volcanic bed within the Queen City sand. Five samples from the middle Cretaceous (Trinity) sands have been included for comparison. Observe the upward increase in idiomorphism, due to increasing volcanism; the biggest change occurs between the Sparta and Yegua sands.

population is somewhat increased by abrasion in the final depositional environment.

SIGNIFICANCE OF IDIOMORPHIC ZIRCONS

The primary aim of this study was to determine if the presence of idiomorphic zircons in large quantities was associated with volcanism. This association now seems proven. Figure 6 is a graph showing the proportion of idiomorphic plus hypidiomorphic zircon (hereinafter termed simply "idiomorphic") as it varies up section. There is an almost perfect correlation of this curve with the degree of volcanism indicated by thin-section evidence (volcanic rock fragments, glass shards, biotite, apatite, plagioclase, and quartz phenocrysts) and field evidence (beds of tuff or bentonite). The Catahoula and Manning formations, which consist almost wholly of vitric tuff and bentonite, have the highest proportion of idiomorphs (45-50 percent); the Carrizo through Sparta section, where evidence of volcanism is least, averages only 30 percent. The major change between a non-volcanic and a richly volcanic section takes place between the Sparta and Yegua sands, based on field and thin section evidence—this is precisely where the biggest increase in idiomorphic zircon occurs. A χ^2 test was run on the proportion of idiomorphic plus hypidiomorphic grains present in the non-volcanic versus the richly volcanic part of the section. This test yielded a χ^2 value of 22.6, which for one degree of freedom produces a P of .00001 — i.e., there is only one chance in 100,000 that this great a difference could arise by chance.

A brief spurt of volcanism is recorded in a six-inch bed of loose, porous sand in the generally non-volcanic Queen City formation, discovered by Diana Riley, University of Texas. Biotite is exceedingly rare throughout the "normal" Queen City but this thin bed contains a flood of subhexagonal biotite and hornblende, and presumably represents an ash fall. This bed has the highest idiomorphic zircon content of any Queen City sample (40 percent), corresponding to the high levels attained only in the Yegua, Caddell or Manning formations of the volcanic-rich part of the section. The evidence appears very convincing that a high idiomorphic zircon content indeed correlates with the activity of volcanism in Texas Tertiary sands.

This association might mean either (1) zircon from volcanic debris is characteristically idiomorphic while zircon from other primary rocks such as granite or schist is less so, or (2) that the volcanic grains were contributed in ash falls directly into the sea or with only brief terrestrial transport so that they remained largely unabraded, while zircon from other sources had to travel more rigorously over much greater distances, or were frequently recycled through older sediments so that more abrasion took place. It is probable that the second factor is more important. Certainly not all the idiomorphic zircon is from a volcanic source, because even the apparently non-volcanic Cretaceous (Comanchean) sands contain 20-25 percent; these were probably derived directly from primary plutonic igneous or metamorphic terranes, which provided a significant part of the Cretaceous detritus in central Texas. Idiomorphic zircon is common in many granites in other areas (Tyler, Marsden, Grout, and Thiel, 1940; Vitanage, 1957).

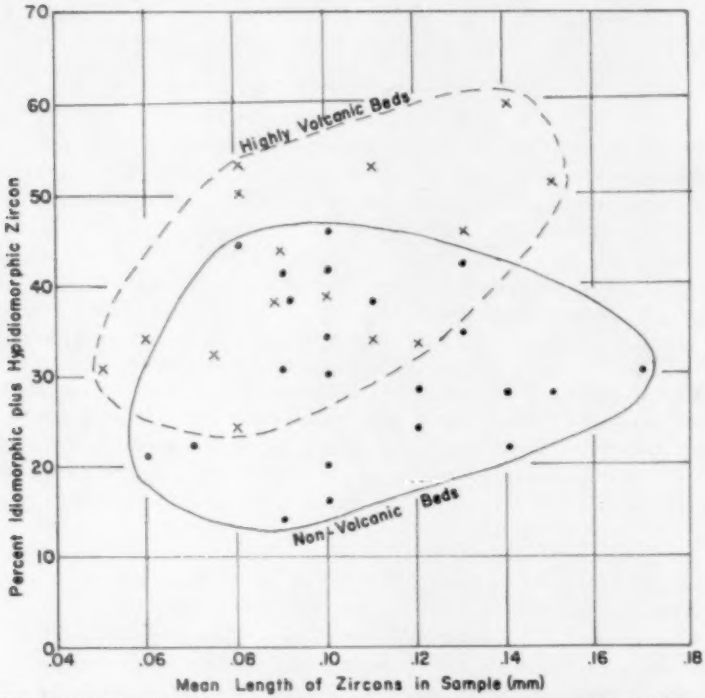

Fig. 7. Scatterplot of mean length of zircon grains versus proportion of idiomorphic plus hypidiomorphic zircon. Each symbol represents a 50-grain count; samples from the non-volcanic part of the section (Simsboro through Sparta) shown by solid dots; those from the highly volcanic section (Yegua through Catahoula) shown by crosses. There is little association of mean size with idiomorphism in grains longer than 0.08 mm; however, grains smaller than this show considerably lower idiomorphism. Size-for-size, samples from the highly volcanic beds show greater idiomorphism.

IDIOMORPHISM VERSUS CRYSTAL SIZE

For each slide, a mean length of the zircon crystals was obtained in order to see if there were any relationship between idiomorphism and crystal size. If there were a relationship, this would prove to be a "muddying" factor in volcanism studies; for example, in a given formation that was being subjected to a constant influx of volcanic material, it is conceivable that a bed containing zircons averaging 0.05 mm long would show a very different proportion of idiomorphs than another bed containing 0.15 mm zircons. Hence one would have to take zircon size into account when predicting volcanic activity. To evaluate this factor, mean zircon size was plotted against percent idiomorphs for all samples (fig. 7). A slight trend appears to exist, in that zircon suites smaller than 0.08 mm tend to be less idiomorphic, but once the average zircon size exceeds 0.08 mm there is very little correlation of grain size with zircon idiomorphism. Thus in studying the relation between idiomorphism and volcanism, it appears that one should not consider slides with zircons averaging

smaller than 0.08 mm as these give misleading low idiomorphism values.

On this graph, zircon counts from the richly volcanic formations are plotted in different symbols from those of the non-volcanic group. It is evident that the volcanic formations plot higher on the idiomorphism scale for each size, with the greatest differences appearing among samples with zircons averaging longer than 0.12 mm.

CONCLUSIONS

It appears that zircon morphology adds another weapon to the arsenal of the petrologist who is trying to correlate formations or decipher paleogeography. Most of the formations studied here yield characteristic zircon assemblages that are functions of source terrane, abrasion in the depositional environment, or volcanic activity. We consider it established that the proportion of idiomorphic zircon is an excellent indicator of volcanism in the lower Tertiary of central Texas, and presume that this relationship will also hold throughout the western Gulf Coastal Plain. Whether it works elsewhere we do not know, although a few spot samples from highly volcanic sediments in other areas also contain abudant idiomorphs, indicating that it may well be a general tendency. Work in the Tertiary elsewhere in Texas ought to establish idiomorphism gradients and should be of great value in locating the source of the volcanic contributions.

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NARROW LINEAR DRUMLINS NEAR VELVA, NORTH DAKOTA*

RICHARD W. LEMKE

ABSTRACT. Remarkably straight and narrow parallel ridges, which are interpreted to be an elongated type of drumlin, trend southeasterly across an area of ground moraine in the vicinity of Velva in north-central North Dakota. The drumlins are paralleled by shallow grooves in the intervening ground moraine; arcuate recessional washboard moraines lie transverse to them. The drumlins are commonly 1 to 3 miles long, 5 to 15 feet high, and have even crestlines. The longest, however, is 13½ miles long and 15 to 30 feet high. The higher and wider drumlins consist mostly of stratified sand and incorporated bodies of till; those less than 5 feet high consist almost entirely of till. It is tentatively postulated that the drumlins were formed in part by glacial erosion of preexisting stratified deposits and in part by till deposition at the base of the ice.

* Publication authorized by the Director, U. S. Geological Survey.

INTRODUCTION

Remarkably straight and narrow parallel ridges trend southeasterly across an area in north-central North Dakota. The purpose of this paper is to discuss the morphology and composition of these ridges, to present reasons why the writer believes that they are an elongated type of drumlin, and to discuss their origin.

The ridges trend southeastward across the Velva area (fig. 1) of about 300 square miles in southern McHenry County. All are confined to the area south of the Souris River valley.

ACKNOWLEDGMENTS

The writer especially wishes to thank Roger B. Colton, Wallace R. Hansen, Joseph H. Hartshorn, and Walter O. Kupsch for their many helpful suggestions concerning the interpretation of the observed features and for critically reading the manuscript.

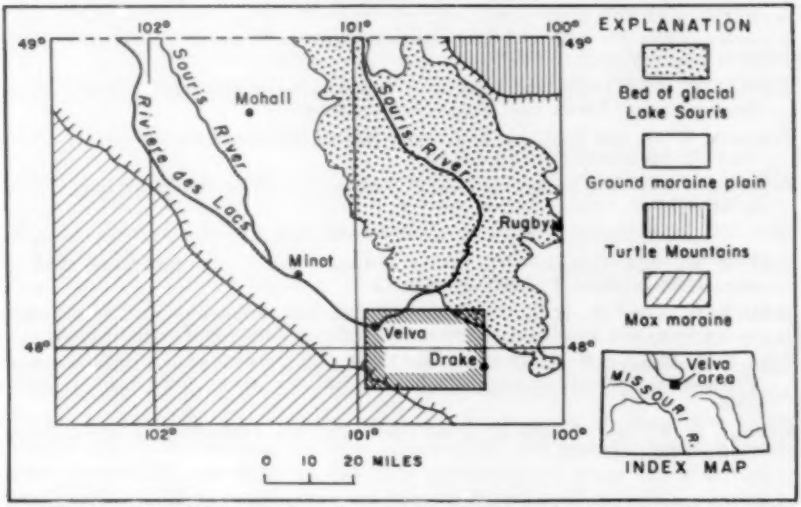

Fig. 1. Generalized map of north-central North Dakota showing major physiographic features in respect to Velva area.

GEOLOGIC SETTING

The marine Cannonball member of the Fort Union formation of Paleocene (Tertiary) age is the uppermost bedrock unit in much of the Velva area. The strata of this member consist chiefly of alternating beds of sand, silt, and silty shale. They crop out in only a few places, mostly along the valley walls of the Souris River. The younger continental Tongue River member of the same formation forms the uppermost bedrock unit in a few places along the western and southern edges of the area and over a large region west and northwest of the Velva area. It is exposed in a few small outcrops and consists mostly of beds of shale, clay, sand, sandstone, siltstone, and lignite.

The linear ridges and related glacial features lie mostly on a ground moraine plain that slopes gently to the northeast and that was formed by a lobe of ice that moved southeastward after the Two Creeks interstadial of the Wisconsin stage. In most places within the Velva area the ground moraine is between 25 and 75 feet thick. The direction of ice advance was transverse to the regional slope except for a slight uphill movement in the southern part of the area. The conspicuous Max moraine (Townsend and Jenke, 1951, p. 842-858), whose surface deposits in this segment of the feature probably belong to the same age of glaciation, lies a few miles southwest of the area of linear ridges; the linear ridges near the Max moraine lie at a slightly higher altitude than do those in the northwest part of the Velva area. The bed of Lake Souris, also formed during this glaciation, lies to the north and northeast of the Velva area. The moraine-covered Turtle Mountains lie northeast of glacial Lake Souris. The relation of these major features to the Velva area is shown in figure 1.

The otherwise nearly flat ground moraine is modified by long linear parallel ridges, shallow linear grooves, low arcuate recessional moraines, wide shallow glacial outwash channels, eskers and kames, and by some other stratified glacial deposits of uncertain origin (fig. 2). The till constituting the ground moraine consists chiefly of silt and sand intermixed with lesser amounts of clay, pebbles, cobbles, and boulders. It is derived in part from the underlying Cannonball member of the Fort Union formation. It therefore contains more sand-size material than the ground moraine to the west and northwest, which is largely derived from the more clayey Tongue River member of the Fort Union formation.

The linear grooves scoured in the ground moraine are parallel to the linear ridges. Some are several miles long but are generally less than 5 feet deep. They are very indistinct in the field but are clearly seen on aerial photos, where their positions are emphasized by the darker color of the vegetation.

Transverse to the linear ridges and grooves are low subparallel arcuate ridges of till and chains of intervening undrained depressions. The chains of undrained depressions are well depicted on the shaded relief topographic map of the Voltaire 7½-minute quadrangle, but the arcuate ridges are mostly too low to be reflected by the 5-foot contours. The transverse ridges are recessional moraines, commonly called washboard moraines (Lawrence and Elson, 1953, p. 95), that may mark cyclic pauses of the ice front during deglaciation.

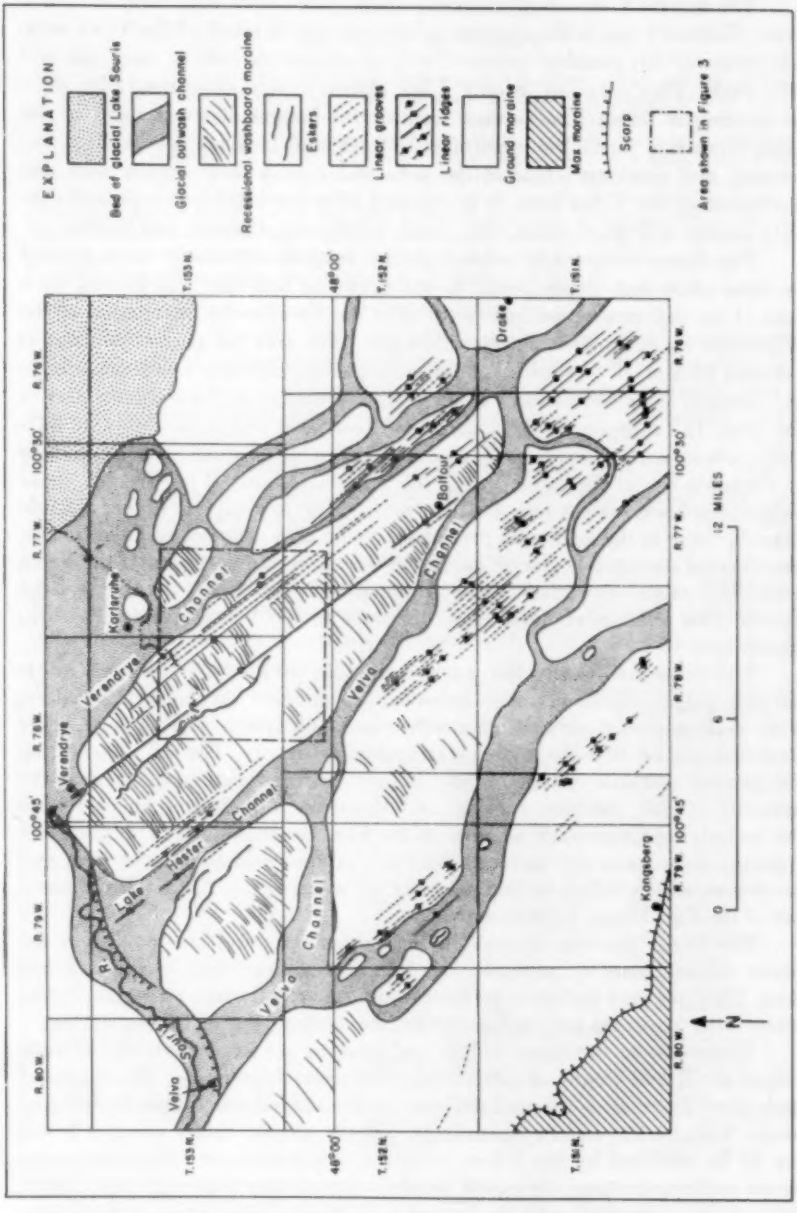

Fig. 2. Generalized geologic and physiographic map of the Velva area.

Most of the transverse ridges consist of segments aligned in arcs; they have uneven crests, generally are less than 10 feet high, and are indiscernible on the ground without the aid of aerial photos or a detailed topographic map. The arcs are concave to the northwest and indicate that the last ice front receded in that direction.

The glacial outwash channels are mostly half a mile to 2 miles wide, 15 to 30 feet deep, and flat-floored. They were formed later than the linear ridges as indicated by the fact that the ridges were partly eroded away by meltwater flowing down the channels. The Velva, Lake Hester, and Verendrye channels were cut by meltwater diverted out of the valley of the Souris River when the downstream part of the valley was blocked by ice and the water was forced to flow southeastward. They marked successive ice-marginal positions during shrinkage of the southwest side of the ice lobe, the Velva channel being formed first.

MORPHOLOGICAL CHARACTERISTICS OF THE LINEAR RIDGES

Most of the linear ridges are 1 to 3 miles long, 150 to 200 feet wide at their base, 5 to 15 feet high, steep-sided and sharp-crested. The average length to width ratio is about 60 to 1. They are only slightly asymmetrical in longi-

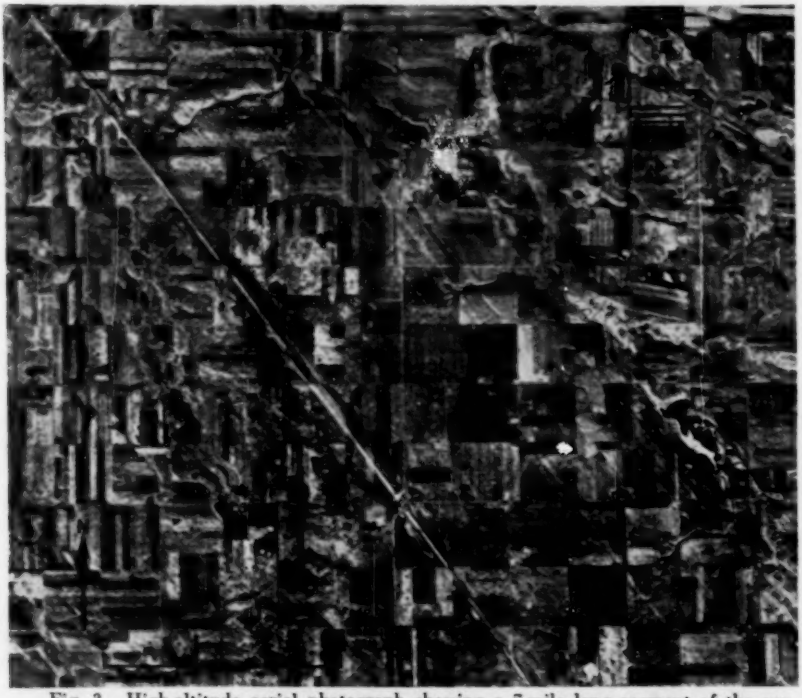

Fig. 3. High-altitude aerial photograph showing a 7-mile long segment of the conspicuously linear ridge that extends southeastward from Verendrye and also smaller linear ridges and linear grooves to the northeast.

tudinal profile (highest at their northwest ends) and are symmetrical in transverse profile. Those less than 5 feet high are scarcely discernible in the field. However, even the lowest ridges are plainly visible on aerial photographs where their crests show as light-gray lines. In cross section, the lowest ridges have gentle, sloping sides and tops whereas the higher ones are proportionately narrower and more conspicuous.

The longest and most conspicuous ridge extends southeast for a distance of 13½ miles (see figs. 2 and 3) from the south valley wall of the Souris River at the town of Verendrye. It is breached in two places near Verendrye by gaps eroded by meltwater flowing down the Verendrye outwash channel and in one place by a small tunnel valley (Flint, 1947, p. 160) cut by a subglacial stream that formed an esker along one segment of its course. The ridge decreases in height from about 50 feet near its northwest terminus to less than 5 feet near its southeast end. Throughout much of its length it is about 30 feet high, is even and sharp crested, and has a remarkable symmetrical cross profile. It resembles nothing so much as a large railroad or highway grade. Its average base width is about 300 feet, which makes its length to width ratio about 240 to 1. At the southeast end of the ridge and slightly en echelon to it is a second ridge, about 3 miles long, which extends to a point about 1½ miles southeast of Balfour.

The highest and widest ridge occupies the middle part of the head of Lake Hester outwash channel. It is about 60 feet high at its northwest end. Southeastward, it diminishes to about 30 feet in height in a distance of slightly less than one mile where it is breached by meltwater that flowed down the outwash channel. A fairly even-crested segment southeast of the breach is nearly 2 miles long. This linear ridge and several others are well shown on the shaded relief topographic map of the Voltaire 7½-minute quadrangle.

Most of the linear ridges trend about S. 50° E. A few bend slightly along their courses. An individual ridge commonly tapers to extinction along its southeast trend, but in many places another ridge is aligned with it or is parallel to it in en echelon fashion. As denoted by figure 2, the smaller linear ridges tend to occur in localized swarms in areas where the linear grooves are also most abundant. However, the larger ridges tend to be more isolated.

COMPOSITION OF THE LINEAR RIDGES

The larger ridges consist predominantly of stratified sand and incorporated irregular bodies of till. Till also commonly forms the flanks of the ridges and in some segments the upper 2 or 3 feet of their crests. The incorporated till bodies and the till forming the surface deposits range from compact till similar to the adjacent ground moraine to a predominantly sandy till that contains just enough clay to bind it. Some of the till has an indistinct fissility or platy structure.

Ridges less than 5 feet high consist chiefly of till. This till appears to be identical to the till composing the adjacent ground moraine.

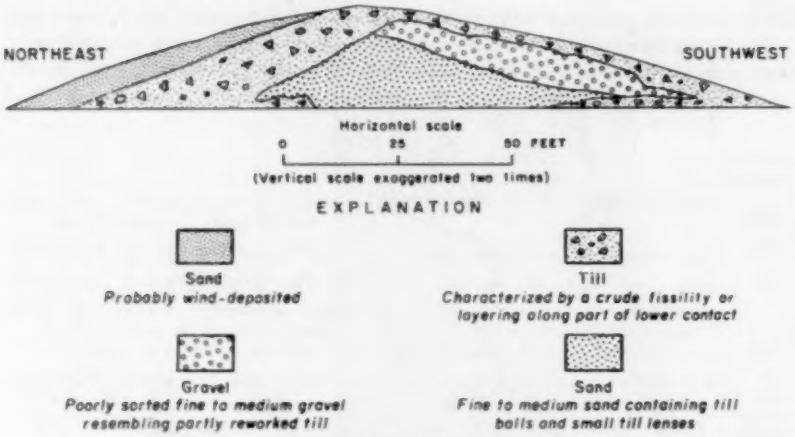

Fig. 4. Generalized cross section of linear ridge exposed in southeast side of roadcut at Balfour in the SW¼ sec. 28, T. 152 N., R. 77 W.

The ridges probably vary considerably in composition and internal structure along different segments of their lengths. As shown in figure 4, the southeast side of a roadcut across a ridge near Balfour exposes a central body of fine gravel and a body of sand flanked and partly underlain by till. The northwest side of the roadcut, although poorly exposed, appears to consist almost entirely of sand. Numerous auger holes drilled in the larger ridges also indicate considerable variability in composition and structure.

No evidence has been noted that the ridges were deposited in the lee of bedrock knobs, boulders, or pre-existing masses of frozen till or blocks of ice or that they have bedrock cores. An auger hole drilled in the northwest end of the large ridge at the head of Lake Hester outwash channel shows that the bedrock surface is no higher under the ridge than under the surrounding terrain and that the composition is typical to that found throughout the lengths of the other larger ridges. Similar stratigraphic relations are believed to exist for the other ridges except for one segment of a ridge in the southeast part of the Velva area where bedrock (considerably tilted and deformed) exposed in a roadcut forms the core of the ridge.

Figure 5 is a cross section based on 11 auger holes drilled across the conspicuous ridge that extends between Verendrye and Balfour. The holes were drilled on that part of the ridge that is about 5 miles southeast of Verendrye. It could not be determined from the drilling whether there is a sharp contact between an upper buff sand and a lower grayish-brown sand as seen in the cross section or whether there is a gradational color change between an upper oxidized sand and a lower unoxidized sand. An additional auger hole was drilled on the same ridge 5 miles farther to the southeast where the ridge is 12 feet high. The top of the hole was in a roadcut 7 feet below the crest of the ridge. Lenticular beds of sand and fine gravel and a few lenses of till, 3 to 5 feet long and 1 to 3 feet wide, are exposed in the roadcut between

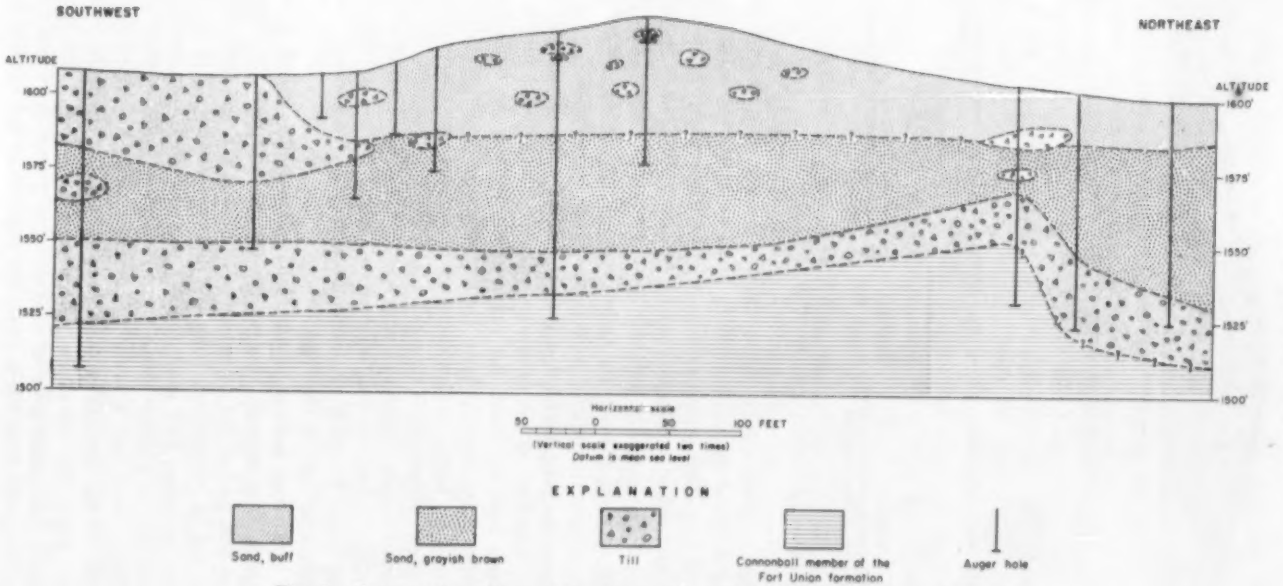

Fig. 5. Cross section of long ridge that trends southeast from Verendrye. Cross section is about 5 miles southeast of Verendrye in the SE¼ sec. 15, T. 153 N., R. 73 W.

the crest of the ridge and the top of the hole. The long axes of most of the stratified and unstratified lenses appear to be parallel to the trend of the ridge. The following material, which is fairly typical of material found in numerous other holes augered in the larger ridges, was penetrated in the auger hole.

Material

	Thickness (in feet)
Sand, tan, medium to coarse, clean; pebbles as much as one-fourth inch long; water starting at about 15 feet	30.0
Till, bluish gray; unoxidized	2.0
Sand, gray, medium-grained; some till lenses	4.5
Sand, grayish blue; believed to consist largely of material derived from Cannonball member of the Fort Union formation (Paleocene) and may include weathered top of that member; hard layer encountered at 65 feet	28.5
Cannonball member of Fort Union formation, green sand and chocolate-brown silty clay	5.0
Total	70.0

INTERPRETATION OF THE LINEAR RIDGES AS DRUMLINS

It is believed that the linear ridges are long, narrow drumlins that have been formed on a till plain during the southeast advance of the last ice lobe to occupy the area.

Before the internal composition, depth to bedrock, relation to transverse arcuate recessional moraines, and other stratigraphic relations of the ridges were well known, the writer and others favored the hypothesis that the linear ridges were crevasse fillings (Lemke, Jensen, Smith, 1951, p. 1460; Lemke, 1953, Geol. Quad. maps). The interpretation that the ridges are long, narrow drumlins was first proposed by the writer in 1954 (p. 1380). The following considerations support this later interpretation and negate the former hypothesis:

1. The ridges conform to the distinguishing characteristics of drumlins in being oriented in the direction of ice movement as indicated by the fact that they lie transverse to the washboard moraines. Also, in conformity to most drumlins, their stoss ends are higher than their tapering lee ends. Crevasse fillings are characteristically developed in fields in which two or more directions of ridge orientation prevail. Moreover, they do not, as a group, possess the streamlined form displayed by these ridges.

2. The pronounced parallelism of the ridges and their extreme linearity, maintained over distances of several miles, can be explained for drumlin formation. The writer, however, knows of no mechanism for fracturing glacier ice that would produce such long, parallel crevases as would be necessary to form crevasse fillings.

3. Linear grooves and linear ridges are commonly associated features on a drumlinized till plain but the formation of linear grooves cannot be explained as associated features in the forming of crevasse fillings.

4. As will be discussed later, partly stratified drumlins are not rare; stratified materials have been noted in many drumlins in both North America and Europe. Flint (1947, p. 124) notes that the remarkably long, narrow

forms in northern Saskatchewan and Manitoba consist of a very sandy till. Whether parts of these features consist of stratified material is not known. It would be difficult to explain by the crevasse hypothesis why the cores of some ridge segments consist of stratified material while their flanks and crests commonly consist of till.

The question may be raised whether such long, narrow ridges should be called drumlins or whether they should more appropriately be designated as drumlinoids, drumlinoidal features, or as ispatinows (Flint, 1947, p. 124). However, the writer believes that the term drumlin should be used because the features conform to the general definition of a drumlin in consisting of glacial drift that has been shaped into a streamlined form by moving basal glacier ice. As pointed out by Flint (1947, p. 121-124), the ideal drumlin form is half-ellipsoidal in shape but that many vary from this ideal and grade all the way from rounded drift hills to remarkably long, narrow forms, present in northern Saskatchewan and Manitoba, that are nearly identical in form to those in the Velva area. As will be discussed later, the long, narrow variety is the common form in many localities not previously discussed in the literature. In some of these localities, all gradations in shape exist between these features and the better known ellipsoidal-shaped drumlins.

Indeed, in some of the more classic drumlin areas described in the literature such as those in the Cayuga-Seneca Falls area, New York, aerial photos show that all gradations exist from ellipsoidal-shaped drumlins to long, narrow, tapering forms; many drumlins that were mapped as nearly oval in plan actually have long, tapering lee ends.

SIMILAR DRUMLINS ELSEWHERE

In shape and in their relation to other features, the drumlins in the Velva area resemble, in one or more ways, features described by Tyrrell (1896, p. 23d), Gilbert (1904, p. 77-78), Alden (1905, p. 17), Grant and Higgins (1913, p. 66), Flint (1947, p. 124-125), Armstrong and Tipper (1948, p. 287-293), Smith (1948, p. 503-513), Hage (1945, p. 24-26), Deane (1950, p. 12-13), Chapman and Putman (1951, p. 166 and p. 199), Dyson (1952, p. 204-211), Hoppe and Schytt (1953, p. 105-115), Colton and Lemke (1955, p. 1673), and others. In addition, similar features have been observed elsewhere by the writer.

One of the earliest descriptions of linear ridges and grooves in ground moraine is furnished by Gilbert (1904, p. 77-78) in describing an area immediately in front of the Columbia Glacier in Alaska. He noted that the ground moraine was elaborately fluted in the direction of ice motion, the corrugations having a vertical magnitude of several feet. In one place he observed that a large boulder in the underlying drift had impressed its form on the ice and preserved a ridge of drift in its lee of the same cross section as the boulder.

Ridges similar to those described by Gilbert were seen by the writer immediately in front of Athabasca and Saskatchewan outlet glaciers of the Columbia Icefield in Canada. These ridges are as much as 1,000 feet long, 1 to 3 feet high, and are remarkably straight, parallel and even crested; they

consist of very compact till, and they trend in the direction of ice movement. Small transverse recessional moraines and deposits brought to the surface of the ice along shear planes are superimposed on the linear ridges. The stoss ends of some linear ridges were buried under the ice. Exposed stoss ends commonly begin in the lee of boulders only slightly higher and wider than the ridges themselves. Nearly every such boulder is conspicuously striated parallel to the ridges.

Dyson (1952, p. 204-211) described linear ridges of till parallel to the direction of ice movement in front of Grinnell and Sperry glaciers in Glacier National Park. The writer observed those in front of Sperry Glacier. Most of the ridges are 2 to 12 feet wide, 1 to 3 feet high, and 500 to 600 feet long. The exposed stoss ends of most ridges terminate against boulders whose top surfaces are worn smooth and striated parallel to the trend of the ridges.

Hoppe and Schytt (1953, p. 105-115) describe ridged ground-moraine surfaces in front of three glaciers in Iceland as well as in front of several northern Scandinavian glaciers. Many of the ridges of the ground moraine surface were in the lee of large partly embedded boulders, but in many places the boulders were absent.

The writer examined elongated drumlins and associated parallel grooves near the southwest edge of the Bearpaw Mountains in northern Montana. Here the ridges are in the lee of igneous dikes. However, examination of nearby areas of drumlinized till showed no evidence that bedrock furnished a buttress for the accumulation of the deposits.

Conspicuous long, narrow, parallel, southeast-trending drumlins and grooves were first noted on aerial photos by Roger B. Colton of the U. S. Geological Survey, in Blaine County, Mont., about 7 miles southeast of Harlem. They were later examined by the writer. In general appearance they are nearly identical to those southeast of Velva except that they bend slightly along their courses suggesting a slight change in direction of ice advance. No evidence was noted that the drumlins had been formed in the lee of any bedrock obstruction.

Aerial photos (furnished by Walter O. Kupsch of the University of Saskatchewan) show that long, parallel drumlins and associated grooves modify the ground-moraine plain near North Battleford, Saskatchewan. Some of these drumlins and grooves are several miles long and are about the same height, shape, and width as the drumlins and grooves of the Velva area. Walter O. Kupsch and Earl A. Christiansen of the University of Saskatchewan have noted (oral communications) linear ridges of probable similar origin in southern Saskatchewan and John A. Elson of McGill University has mapped this type of feature in southern Manitoba (personal communication).

Colton and Lemke (1955, p. 1673) noted long, narrow drumlins in several other areas in North Dakota in addition to the drumlins in the Velva area. All these areas are east of the Velva area.

Armstrong and Tipper (1948, p. 287-293) described widespread, parallel, elongated, drumlinlike ridges of till and intervening rectilinear depressions or grooves in north-central British Columbia. The ridges range in shape from some that are nearly circular, to those that are oval, and to those which are

5 to 10 times as long as they are wide. Most are 50 to 75 feet high and the stoss ends are generally steeper, higher, and wider than the lee ends. Commonly they are aligned in groups 5 to 6 miles long, separated laterally by parallel trough-like valleys 50 to 150 feet deep. Rock drumlins are interspersed among the till drumlins, but the two types nowhere overlap.

Tyrrell (1896, p. 23d) described steep, narrow ridges in the Cree Lake area in northern Saskatchewan that are parallel to the direction of glaciation and have a characteristic drumlinlike form; they consist of loose unassorted material. He called these streamlined features ispatinows, the Cree word for conspicuous hill. The writer examined aerial photos (furnished by W. O. Kupsch) of part of the Cree Lake area and noted that all gradations exist in form between the more classic oval-shaped drumlins and narrow linear ridges as much as 2 miles long.

The writer observed a well-drumlinized area in southwestern Glacier County, Mont., between Little Badger Creek and Two Medicine Creek. The drumlins here range in shape from the classic type of drumlin, which is oval in plan and asymmetrical in longitudinal profile, to longitudinally nearly symmetrical narrow ridges about 20 feet high and more than a mile long that are identical in appearance to those southeast of Velva. Their crests and flanks consist mostly of till; their internal composition is not known.

ORIGIN OF DRUMLINS IN THE VELVA AREA

From the above descriptions and from known occurrences of similar features elsewhere, it is evident that long, narrow drumlins are widespread. Flint (1947, p. 125) notes that this type of drumlin commonly consists of sandy till, which is analogous to the Velva drumlins. However, many of the drumlins from the Velva area contain stratified material and therefore, so far as is known, differ in this respect from long, narrow drumlins described elsewhere. However, the composition of most of the long drumlins elsewhere is not well known, and it is possible that some of them contain stratified material.

Dyson (1952, p. 204-211) believes that the linear ridges of till in front of Grinnell and Sperry glaciers were formed by partial filling of tunnels at the base of the ice in the lee of boulders; the till was squeezed into these tunnels from unfrozen drift at the base of the ice. That such tunnels do exist at the base of the ice in the lee of boulders or bedrock obstructions was demonstrated by Carol (1947, p. 57-59), who penetrated such tunnels for distances of more than 150 feet. Hoppe and Schytt (1953, p. 105-115) also postulate that till was pressed into tunnels in the lee of boulders to form linear ridges in front of glaciers in Iceland. However, unlike Dyson's interpretation, they found that, in uncovering some of the ridges from beneath the ice by trenching, the ridges had a high ice content of frozen meltwater, and that the till under the glacier was frozen throughout the year to a point some distance back under the ice. They postulate that the weight of the ice was sufficient so that the water-soaked frozen ground moraine reached the pressure-melting point and was pressed into the tunnels. The explanations for forming linear ridges offered by Dyson (1952) and also by Hoppe and

Schytt (1953) do not apply to the drumlins in the Velva area because the Velva drumlins were not formed in the lee of a boulder or bedrock obstruction. Moreover, it would not be possible for ridges formed in the lee of such obstructions to contain stratified material solely by squeezing material up into the tunnels. However, one might consider the possibility that the stratified material was deposited by a meltwater stream using the tunned at the same time that the till was being squeezed into the opening from the underlying ground moraine and also from debris in the basal till-rich glacier ice forming the walls and top of the tunnel.

Armstrong and Tipper (1948, p. 287-293) believe that a knob of frozen till, rather than boulders or bedrock obstructions, formed buttresses to the advancing ice and that till was deposited in the lee of these obstructions to form the drumlins in north-central British Columbia. Whether similar knobs of frozen till constituted the obstructions for forming the drumlins in the Velva area cannot be determined. However, no knobs of till now are present at the stoss ends of the drumlins to support such an assumption. It might be postulated that overriding of isolated blocks of ice in front of the glacier would offer suitable obstructions. Again there is no evidence, such as the presence of kettles at the stoss ends of the ridges, to support such an assumption.

Partly stratified drumlins are not rare. As pointed out by Gravenor (1953, p. 676), stratified materials have been noted in many drumlins in both North America and Europe. Most such drumlins, however, are markedly asymmetrical in longitudinal profile, oval in plan, and considerably steeper and wider at the stoss end than at the gently sloping and tapering lee end. Their lengths rarely exceed five times their widths and generally are considerably less.

Kupsch (1955, p. 327) believes that some oval-shaped stratified drumlins near Dollard, Saskatchewan, were originally kames that were overriden by ice and shaped into a streamlined form. He refers to these features as "drumlinized kames." A few drumlins of this type are in Barnes County, N. Dak., in areas a few miles northeast and southeast of Valley City. Their cores consist of stratified deposits and their crests and flanks consist of till. They trend southwest and appear to have been formed by ice advancing over stratified ice-contact deposits and plastering till on top.

Gravenor (1953, p. 674-681) believes that drumlins can be formed as follows:

"(1) Masses of till and stratified materials would be deposited at the front of an advancing glacier if there was a temporary halt during the ice advance.

"(2) Ice riding over this drift would erode and shape it and thereby produce drumlins."

According to this theory, both erosion and shaping of existing deposits occurs. Hence any pre-existing thin stratified deposits that might be present at the surface would be preserved only where these deposits were moulded into ridges.

The erosion and shaping theory for the origin of drumlins, as pointed out by Gravenor, offers an explanation for the presence of stratified material in typically shaped drumlins. Whether the interpretation is applicable to the origin of the long, narrow drumlins in the Velva area, however, is open to question. Assuming that there was stratified drift in front of the last advancing ice sheet to cover the Velva area, the question arises whether this stratification along the trend of such a long and narrow ridge would be preserved during the time the ridge was being formed in part by moulding and in part by erosion. Moreover, this interpretation does not account for till on the flanks and crests of most of the larger ridges. One might, however, postulate that the till was deposited from the basal overriding ice—certainly a likely possibility. Also unexplained is how the till masses and the stratified deposits of the ridges could be so intertongued and otherwise intimately related without destroying the stratification of the bedded deposits unless it is interpreted that the bodies of till were incorporated during original deposition of the stratified material. Also left unexplained is why the small ridges consist almost entirely of till if there was merely erosion and shaping of surface stratified deposits.

The origin of the linear grooves parallel to the linear ridges also is not readily explained by Gravenor's interpretation. Although grooving could take place simultaneously with the removal of the rest of the material between the ridges, the question remains why erosion would take place down to one level to produce the present nearly level ground moraine surface and then, here and there, the surface would be further scoured to a rather uniform depth to produce the grooves.

In summary, it is believed that the linear ridges are drumlins formed during a southeast advance of the last ice sheet to occupy the Velva area. Although leaving some questions unanswered, the writer tentatively favors the hypothesis that the larger partly stratified drumlins were formed by glacier ice advancing over stratified deposits, eroding and shaping the deposits into ridges and then plastering a layer of till over the deposits. Whether the smaller drumlins consisting entirely of till were formed by ice erosion, by a plastering-on process, or by till being pressed up into linear tunnels at the base of the ice, is not clear.

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PERIGLACIAL LOESS, I MORPHOLOGICAL PROPERTIES*

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ABSTRACT. The physical and mineral properties of periglacial loess and alluvium of the Susquehanna River Valley, Pennsylvania and a section of the Laurentians, province of Quebec, are compared to find means to correlate the field and laboratory observations of their morphological properties.

Because of the common occurrence of similar particle-size distribution and porosity in loess and in alluvium, these properties must be combined to be used as a reliable criterion of loess. Sand grains are more frosted and angular in loess than in alluvium, and in Pennsylvania loess than in Quebec loess in agreement with the pronounced vertical parting of Pennsylvania loess. There is a disparity between the minerals found in loess and underlying material in Pennsylvania. The minerals found in the sand fraction of loess vary with the minerals in the source material. The color of loess is due to surface coatings of the particles and it is possible that other than buff or tan colored loess occurs in nature.

INTRODUCTION

During field investigations to identify periglacial loess in the Susquehanna River Valley, Pennsylvania and a section of the Laurentians, province of Quebec, Canada, the authors encountered difficulty in differentiating loess from fine alluvium in regions where glacial and/or periglacial climate once prevailed.

Since the appearance of the word "loess" in geological terminology, numerous characteristics, such as high degree of size sorting, large porosity, angularity and frostiness of grains, predominance of light fraction minerals, tan color and mineral disparity from the underlying material, have long been considered as exclusive or characteristic properties of loess (Colby et al., 1953; Doeglas, 1948; Smith, 1942; Zeuner, 1949). However, there has been some speculation (Fisk, 1947; Péwé, 1950; Russell, 1944) about the occurrence in whole or in part of these properties in material of alluvial origin.

The purpose of this study is to compare the physical and mineral properties of periglacial loess and alluvium[1] and to discover whether there is a significant difference in these properties.

MATERIALS AND METHODS

Areas studied.—In Pennsylvania, loess and loess-like alluvial material were examined and mapped from Middletown on the main Susquehanna River to Williamsport and Towanda on the West and North Branches respectively. In the Laurentians, province of Quebec, a similar investigation was conducted in the area located between St. Jerome and Fort Coulonges, Mont Laurier and the Ottawa River, a distance of 140 miles east to west and 80 miles north to south.

Identification of loess in the field was based mainly on visual observation and on interpretation of the morphological properties herein described, in conjunction with physiographic evidence and properties that will be discussed

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[1] The word alluvium is here used to include all varieties of water-laid materials in contrast to the wind-laid loess.

in a subsequent article.

Collection and description of samples.—A series of 50 duplicate samples of loess and loess-like alluvium were taken with a core sampler from each of the two investigated areas. The samples generally were collected under similar good drainage conditions.

The samples used in this study are described in table 1. Only samples of unaltered material were selected for this study to eliminate the effects of pedogenic factors upon the particle size distribution and porosity of the material especially. The reaction of all samples was acid.

TABLE 1

Description of Samples

Sample No.	Location	Horizon	pH	Remarks
Loess				
P-12C	Harrisburg, Pa.	C	6.1	120″ thick
P-12C+	Harrisburg, Pa.	C	6.1	120″ thick
P-13	Harrisburg, Pa.	C	4.4	180″ thick
P-13x	Harrisburg, Pa.	C	4.4	180″ thick
P-13Cy	Harrisburg, Pa.	C	4.5	180″ thick
P-13Cy[1]	Harrisburg, Pa.	C	4.3	180″ thick
P-20	Harrisburg, Pa.	C	5.8	180″ thick
P-20A	Harrisburg. Pa.	C	5.7	brown layer
P-14C	1.5m N. Halifax, Pa.	C	5.0	72″-120″ thick
P-16C	2.3m S. Millersburg, Pa.	C	4.5	40″-60″ thick
P-24C	1.5m W. Catawissa, Pa.	C	5.8	36″-48″ thick
P-25B	1.0m W. Bloomsburg, Pa.	C	5.5	48″-72″ thick
C-40	1.5m S. W. Val David, P. Que.	C	5.1	15″-28″ thick on top of mountain overlying till and bedrock
C-37	0.6m S. St. Jovite, P. Que.	C	5.3	20″-30″ thick on top of mountain overlying till
C-34B_3	0.5m S. St. Jovite, P. Que.	C	5.1	18″ thick same mountain as C-37, lower location
C-38	3m S. St. Jovite, P. Que.	C	5.2	24″ thick top of mountain overlying till
C-39	3m S. St. Jovite, P. Que.	C	5.1	28″ thick same mountain as C-38, lower location
Alluvium				
P-10C	0.8m S. E. Coxtown, Pa.	C	4.6	thinly stratified
P-18x	3.0m W. Marysville, Pa.	C	6.1	
P-15B	1.5m N. Halifax, Pa.	D	5.0	lies under loess, sample P-14C
P-16DC	2.3m S. Millersburg	D	4.2	lies under loess, sample P-16C
P-24C+	1.5m W. Catawissa, Pa.	D	4.8	lies under loess, sample P-24C
P-26C	2.5m N. E. Bloomsburg, Pa.	C	5.5	
P-27	1.0m S. Mifflinville, Pa.	C	6.1	
P-21	1.5m N. Selinsgrove, Pa.	C	5.8	
P-C	1.7m N. Selinsgrove, Pa.	C	5.1	
P-E	1.2m N. W. Montandon, Pa.	C	5.6	
C-42C	4m S. E. Brebeuf, P. Que.	C	5.8	high flood plain material
C-29C	Venosta, P. Que.	C	4.7	Varved, lacustrine silt
C-30	2m N. Maniwaki, P. Que.	C	5.7	high flood plain material

C-31	2m N. Bois Franc, P. Que.	C	5.2	high flood plain material
C-33	3m N. Ferme Neuve, P. Que.	C	5.5	high flood plain material
C-25B_3	Wyman, P. Que.	B_3	5.3	Varved lacustrine material
C-13	2.2m N. E. Bryson, P. Que.	B_3	5.5	Varved lacustrine material
C-26C	Calumet Village, P. Que.	C	6.7	Varved lacustrine material
C-35C	1.5m S. W. St. Jovite, P. Que.	C	5.7	flood plain, source material of loess sample C-37 and C-34B_3
C-36	0.3m S. St. Jovite, P. Que.	C	5.7	Wind reworked glacial outwash, same mountain as C-34B_3, lower location
C-43C	2m N. E. Norway Bay, P. Que.	C	6.5	

Preparation of samples.—The air-dried samples were sieved through a 2 mm sieve, the few hard aggregates being crushed with a rubber-tipped pestle. The gravel fraction present in some water-laid materials was examined, visually evaluated and then discarded.

Mechanical analysis.—The pre-treatment for the destruction of organic matter was omitted as organic matter was in negligible amount. The hydrometer method (Bouyoucos, 1936) was used in conjunction with a nomographic chart (Casagrande and Fadum, 1940). The sands were fractionated, prior to the hydrometer determinations, with a nest of sieves, No. 18, 35, 60, 140 and 270. Hydrometer readings were taken at intervals of 40 seconds, 1, 2, 5, 15, 30 and 60 minutes, 2, 6, and 24 hours to determine the approximate size of falling particles.

Absolute specific gravity.—The true density was determined for the fine fraction and for the whole soil (Lyon and Buckmann, 1950). The data from the fine fraction were used to designate the particle size at a specific time on Casagrande's nomographic chart, whereas the data from the whole soil were used to calculate the bulk density values of the samples.

Bulk density.—The bulk density was determined by the ratio of the total weight of soil, expressed on a moisture free basis, to the total volume of the soil in the cores.

Hygroscopic moisture.—The loss of weight of a measured quantity of soil dried to constant weight at 110°C was expressed as percent of moisture-free soil.

Estimation and identification of minerals in the sand fraction.—Samples were treated with a 1-1 solution of boiling hydrochloric acid for thirty minutes to remove iron stains (Cailleux, 1942).

Two sand fractions 0.25-0.10 mm and 0.10-0.05 mm were separated into two specific gravity sub-fractions, above and below 2.70 (Marshall and Jeffries, 1945).

A staining method (Reeder and McAllister, 1957) was used to identify the plagioclase and orthoclase in the fine sand fraction. Minerals of the heavy sub-fraction were identified by means of a petrographic microscope.

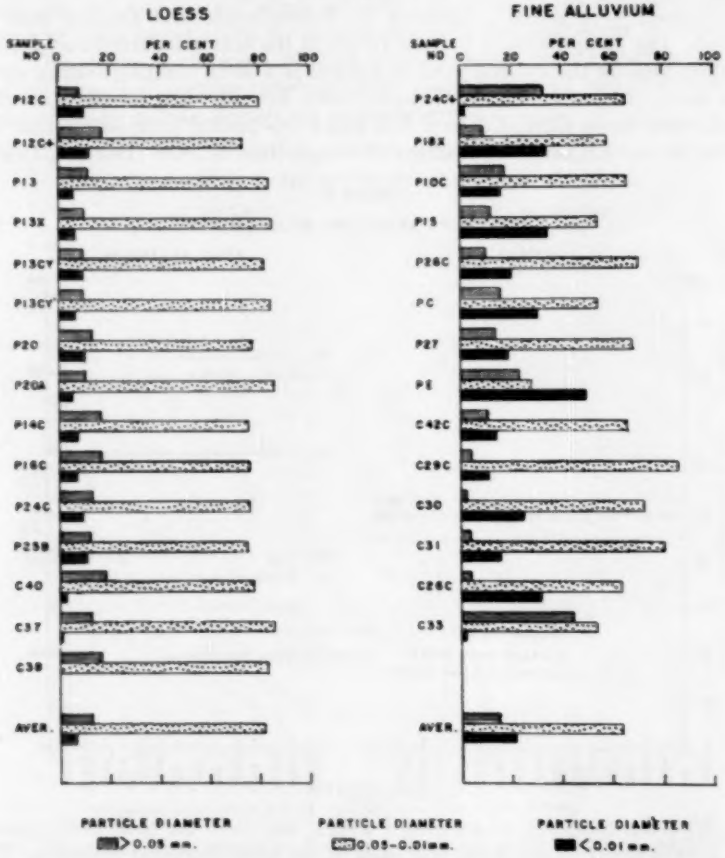
RESULTS AND DISCUSSIONS

Texture and Porosity.—The mechanical composition and porosity of loess and loess-like alluvium are compared in graph 1 and 2 respectively. The

different size limits as used in the graphs follow the size range used by Doeglas (1948). According to Doeglas, loess normally consists of a maximum of 5 percent of particles larger than 0.05 mm and a minimum of 65 percent of particles within the range of 0.01-0.05 mm.

GRAPH 1

MECHANICAL COMPOSITION

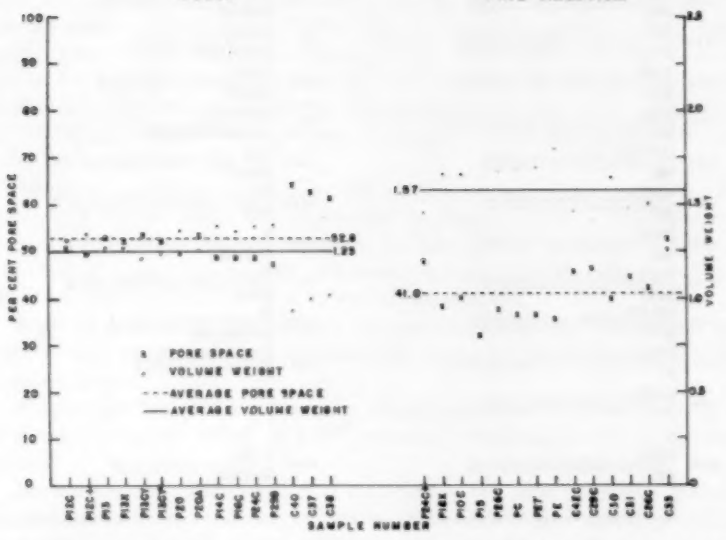


The average figures in graph 1 indicate that loess in general contains a larger percentage of particles within the size range of 0.05-0.01 mm than fine alluvium whereas the lowest average for particles with a diameter less than 0.01 mm occurs in loess. In spite of this apparently conclusive support for Doeglas' normal texture, there is a range of variations in the mechanical composition of both deposits, loess and fine alluvium, which limits the usefulness of size sorting as a criterion of loess identification. The range of variations in the percentage of particles that fall within the specified size limits,

>0.05 mm, 0.05-0.01 mm and <0.01 mm are 9.2-18.5, 73.0-86.0 and 0-11.0 for loess, 2.2-44.8, 28.0-85.9 and 1.8-49.0 for alluvium respectively. These ranges compare significantly with those obtained by Smith (1942) for the mechanical composition of loess, namely 0.5-24.8, 65.0-87.3 and 4.8-32.0 percent respectively. The loess identified in Pennsylvania and Quebec had a more uniform yet coarser texture than that described by Smith (1942). This comparatively coarse texture may indicate that the material making up the loess was carried over shorter distances in Pennsylvania and Quebec than in Illinois. The comparatively narrow range in the textural variations of loess substantiates the theory that wind in general is a more exacting sorting agent than water (Bagnold, 1941). It is noteworthy, however, that out of 14 samples of alluvium three, C29C, C30 and C31 had a mechanical composition that fell within the normal range of mechanical composition of loess (Doeglas, 1948).

GRAPH 2

TOTAL PORE SPACE AND VOLUME WEIGHT



With the exception of samples P-24C+ and C-33 the total pore volume was always lower in the water-laid than in the wind-deposited materials. The trend in bulk density is the reverse. The large total pore volume encountered in the loess from the Laurentians appears to be associated with its relative shallowness in the field, less than 30 inches, in comparison with 180 inches in Pennsylvania. In the latter area the thickness of material probably exerted sufficient pressure to decrease the total pore volume. Although neither total pore volume nor particle size distribution can characterize loess if considered individually, their combined results seem to differentiate loess from loess-like alluvium. In other words loess-like particle size distribution and large total

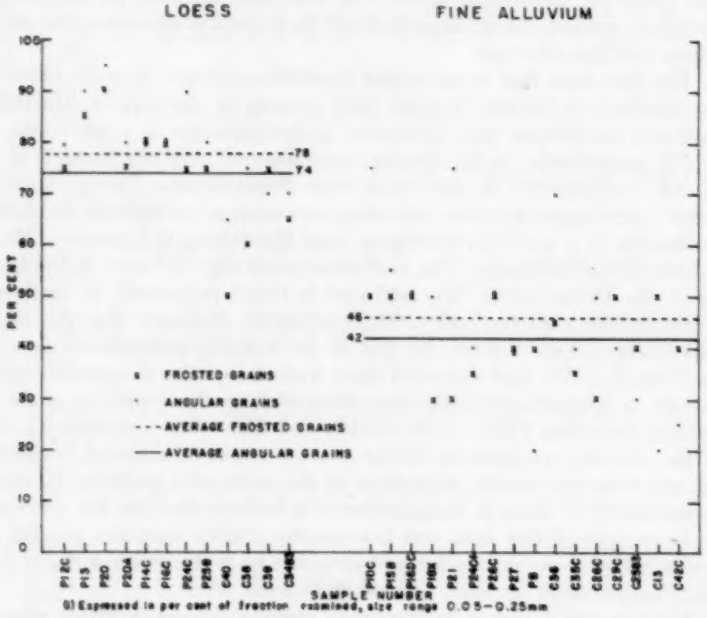
pore volume seldom parallel each other in loess-like alluvium. The statistical significance and mathematical limitations of these figures will be discussed in a subsequent article.

Frostiness and Angularity of Sand Grains.—It has been suggested in the literature (Colby et al, 1953; Fuller, 1922; Smith and Fraser, 1935) that grains of wind-borne materials generally have an angular shape and a rough surface because of insufficient wear and that individual grains, resting on their angular and frosted edges, probably are responsible for the vertically standing walls found along the erosion channels that may occur in loess.

The grains of the sand fraction were examined with a microscope and the percentage of angular (predominantly sharp cornered) and frosted grains determined by counts. Results are shown in graph 3.

GRAPH 3

PER CENT ANGULAR AND FROSTED SAND GRAINS(1)



(1) Expressed in per cent of fraction examined, size range 0.05—0.25mm

The calculated averages indicate that the percentage of angular and frosted grains is larger in loess than in fine alluvium. The data show that out of 16 samples of alluvium only three samples have 70 percent or more of angular grains whereas out of 13 samples of loess, 12 samples have 70 percent or more of angular grains. Again in the alluvium, the maximum proportion of frosted grains is around 50 percent, whereas all but one of the loess samples have well over 50 percent of frosted grains. These comparisons support the statements (Clarke, 1916; Smith and Fraser, 1935) that loess grains are characterized by angular shape and rough surface. This property can

further be demonstrated through comparing the results of three paired samples of loess overlying fine alluvium respectively, P-14C—P-15B, P-16C—P-16DC and P-24C—P24C+. Moreover trends of increasing angularity and frostiness occur in the sequence of samples C-35C, C-36 and C-34B_3 which are geographically associated and come respectively from the alluvial source material, an intermediate wind reworked glacial outwash and the wind-deposited material.

In general the sand particles of loess are more angular and frosted in Pennsylvania than in Quebec; likewise the loess parts vertically better in Pennsylvania than in Quebec.

Mineral Composition of the Sand Fraction.—It has been postulated in the past (Clarke, 1916; Smith and Fraser, 1935; Udden, 1894) that loess is predominantly made up of lightly colored minerals of low specific gravity and that this could be responsible for its common tan or buff color. In table 2 are listed the percentage of the very fine sand fraction, its percentage in each of the specific gravity separates and its mineral composition, for samples of loess and fine alluvium.

The data show that hematite and limonite are absent from the Laurentian loess whereas the content reaches 30.3 percent in the loess at Harrisburg. Conversely hornblende and tourmaline percentages are at a maximum, 19.2 and 7.0, respectively, in the Quebec loess whereas they are lower than 10.4 and 1.8, respectively, in the loess from Pennsylvania. The hematite and limonite percentages decrease and the percentages of hornblende and tourmaline increase in a northerly direction, from Harrisburg to Catawissa, Pa. and the Laurentians in Quebec. The orthoclase percentage is larger in the Quebec than in the Pennsylvania loess and coal is found exclusively in the loess of the Susquehanna River Valley. This variability indicates that the mineral composition of loess depends on that of the material available in the source area (Chepil, 1957) and that wind has a limited capacity to segregate specific minerals. A comparison of the data obtained from the sequences of the geographical associates C35C—C36—C34B_3 and C39—C38 (see table 1) shows that the mineral compositions of the alluvial source of material, transitional loess and loess are similar regardless of the mode of deposition. In spite of this similarity in mineral composition it is noteworthy that the increase in the percentages of fine sand and low specific gravity separates parallel each other in both sequences in a leeward direction from the source area, as Smith (1942) suggested in his study of loess in Illinois.

The minerals found in three pairs of samples of loess overlying alluvium, respectively P14C—P15B, P16C—P16DC and P24C—P24C+ indicate that the mineral composition of loess, like other translocated materials, differs at least quantitatively from that of the underlying material, in accordance with the findings of Smith and Fraser (1935). This mineral disparity correlates with the differences previously reported in graph 3 in the angularity and frostiness of the grains of these samples.

Aside from a coincidence in the high percentage of hornblende and the olive cast of sample C29C no correlation appears between color and the presence or predominance of a specific mineral in samples of loess or al-

TABLE 2

Specific Gravity Separates and Mineral Composition of the Very Fine Sand Fraction of Loess and Alluvium

Sample No.	Color[1]	Percent of total sand	Percent of sand fraction with specific gravity		Light Fraction[2]			Heavy Fraction[2]						
					Quartz	Ortho-clase	Plagio-clase	Coal	Horn-blend	Tourma-line	Magne-tite	Hema-tite	Limo-nite	Others
			<2.70	>2.70										
C-35C[3]	lt. yell. brown	20.2	81.7	18.3	92.7	7.2	0.1	0	20.3	3.4	5.0	0	0	71.3
C-36[3]	lt. br. gray	31.8	83.9	16.1	97.8	2.2	0	0	12.8	4.2	7.4	0	0	75.6
C-34B$_2$[4]	lt. yell. brown	63.5	85.9	14.1	95.7	4.3	0	0	17.6	7.0	6.3	0	0	69.1
C-40[4]	br. yellow	89.0	90.0	10.0	96.2	3.7	0.1	0	19.2	5.2	9.4	0	0	66.2
C-39[4]	pale brown	64.0	88.7	11.3	94.8	5.2	0	0	12.8	4.3	4.7	0	0	78.2
C-38[4]	lt. yell. brown	86.0	89.6	10.4	96.4	3.6	0	0	10.0	3.2	3.7	0	0	83.1
C-29C[3]	olive gray	100.0	95.2	4.8	92.6	7.3	0.1	0	35.6	4.6	1.0	0	0	58.8
P-13[4]	pale brown	97.3	94.1	5.9	99.3	0.4	0.2	0.1	0.4	0.4	0.6	19.6	8.0	71.0
P-20[4]	lt. yell. brown	94.0	95.0	5.0	99.4	0.3	0.1	0.2	3.0	1.0	2.1	14.4	10.2	69.3
P-20A[4]	brown	95.7	94.7	5.3	99.5	0	0.1	0.4	1.4	1.8	0.7	24.4	10.4	61.3
P-12C[4]	lt. yell. brown	94.4	92.7	7.3	99.7	0.1	0.1	0.1	1.3	0.4	0.2	30.3	7.4	60.4
P-14C[4]	dk. yell. brown	97.4	95.8	4.2	98.9	0.8	0.2	0.1	6.9	1.4	8.5	26.5	5.2	51.5
P-15B[3]	lt. yell. brown	68.5	97.8	2.2	94.7	5.0	0.1	0.2	9.5	1.4	0.5	2.7	7.6	78.3
P-16C[4]	lt. yell. brown	89.6	96.2	3.8	98.9	0.6	0.3	0.2	6.2	0.4	3.7	10.0	4.2	75.5
P-16DC[3]	dk. brown	81.9	91.8	8.2	98.8	0.2	0	1.0	0	0.3	1.3	23.6	7.0	67.8
P-24C[4]	yell. brown	66.4	97.4	2.6	97.8	1.3	0.6	0.3	10.4	0.6	2.4	0.3	1.5	84.8
P-24C+[3]	yell. brown	60.5	97.0	3.0	95.0	4.1	0.7	0.2	19.5	1.9	4.5	0.4	0	73.7
P-21[3]	yell. brown	59.5	97.0	3.0	95.8	3.2	0.9	0.1	19.0	1.2	3.3	0.4	1.2	74.9

[1] Munsell color chart
[2] Expressed in percent of total grain counts in each fraction
[3] Alluvium
[4] Loess

luvium. It is evident that if the treatment with 1:1 hydrochloric acid bleached the samples, the color of the grains was due to surface coatings. It is therefore possible that other than buff or tan-colored loess occur in nature.

CONCLUSION

The particles in the loess in Pennsylvania and Quebec, although comparable with those in Illinois, are coarser and have a narrower range of variation. Three samples of alluvium have a particle size distribution that falls within the normal range of the mechanical composition of loess, thus indicating that particle distribution, used singly, cannot differentiate loess from alluvium.

Because porosity may be as large in some alluvium as in loess, large pore volume, a remarkably consistent characteristic of loess, remains, if used singly, an unreliable criterion of loess. As large porosity and loess-like particle size distribution rarely occur concurrently in loess-like alluvium, these properties combined appear to provide a justifiable and reliable means of differentiating loess from alluvium.

Microscopic study of the shape and faces of sand grains shows that single particles are more frosted and angular in loess than in alluvium. The large amounts of frosted and angular grains found in the loess of Pennsylvania is associated with a pronounced ability of this loess to retain a vertical wall in the field.

The great variability encountered in the nature of the mineral constituents of loess shows that its mineral composition depends primarily on the mineral composition of the material available in the source area. In spite of this similarity in the mineral composition of loess and source-material there is an increase from source area to loess in the percentage of fine sand and of the low specific gravity separate of the fine sand fraction.

In the Susquehanna River Valley the mineral composition of loess and the underlying alluvium differed at least quantitatively.

The color of the loess is due to surface coatings and seems to be influenced but little by mineral composition. Loess of colors other than buff or tan may, therefore, occur in nature.

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THE DISTRIBUTION OF TEMPERATURE IN FLOWING WELLS

T. BOLDIZSAR

ABSTRACT. The distribution of temperature in the liquid in a flowing well is calculated by an approximate theory based on Jaeger's solution for the conduction of heat from a cylinder in an infinite medium. Satisfactory agreement is found with observations by Birch for a well near Colorado Springs. In principle, analysis of the temperature distribution and rate of flow can provide values of thermal gradient and of heat flow.

INTRODUCTION

A theory of the distribution of temperature in a flowing fluid—water, oil or gas—in drill holes may be of value in geothermal research and in oil production. While the temperature distribution along a stagnant fluid column rapidly approaches that of the rock, appreciable differences may develop when the fluid is in motion. The temperatures in a well in which liquid enters at a single level are calculated below with the aid of simplifying assumptions and compared with observations given by Birch (1947) for a well near Colorado Springs, Colorado. In principle, this analysis, combined with observations of temperature and of fluid flow, can furnish values of mean thermal conductivity and of the local normal heat flow. An approximate theory has been applied by Kunz and Tixier (1955) to the problem of estimating the flow of gas with the aid of temperature surveys in producing wells.

THEORY

We suppose that, at time $t = 0$, liquid enters a vertical cylindrical hole at a single depth H below the surface of the ground, and take the z-axis, positive upwards, along the axis of the hole, with the origin of coordinates at the point of entrance; thus, $z = 0$ corresponds to the level of entrance of the liquid, $z = H$ to the surface of the ground (fig. 1). The initial or undisturbed rock temperature, T_R, is taken to be linear with depth with the uniform

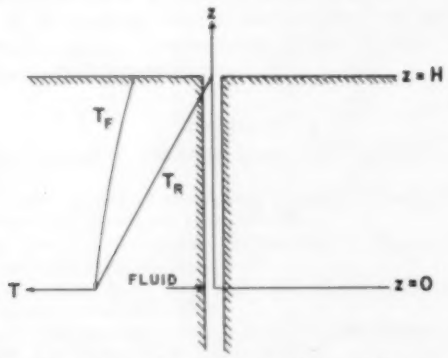

Fig. 1. Diagram showing coordinate system and assumptions regarding temperature. The fluid enters at $z = 0$ with the temperature of the rock at that depth. The temperature T_R of the rock falls off linearly to its surface value, while the temperature of the fluid falls off along some curve, T_F.

gradient σ, so that $T_R(z) = T_R(H) + \sigma(H - z)$. The fluid is assumed to enter with the temperature of the rock at $z = 0$; the rate of flow, supposed independent of time, is V cm³/sec. The conductivity of the rock is K and its diffusivity k; these are taken as uniform. The specific heat of the liquid, per unit volume, is c.

As the liquid moves upward, it loses heat to the cooler rock. The change of liquid temperature T_L is given by $dT_L = -q\,dz/Vc$, where q expresses the rate of loss of heat per unit length of hole. (For the case of liquids, the change of temperature as a result of the change of pressure is negligible, amounting for water to less than 1° C/km; this effect is of major importance for gases). We now suppose that q can be expressed in the form, $q = KF\tau$, where $\tau = (T_L - T_R)$, the difference of temperature between liquid and undisturbed rock, and F is a function which takes into account the change of rock temperature with time as it is heated by the water. F is taken to be a function of time t, diffusivity k, and radius of the hole, a, but not of z. In other words, we shall calculate F for the two-dimensional case, where temperature is independent of z, but employ the result for an approximate calculation of the variation of temperature with z. This procedure is especially appropriate for the geothermal problem where the vertical gradient is small by comparison with the radial gradient near the hole (less than one-half percent in the example considered below).

With these assumptions, the equation for τ becomes

$$dz = d\tau/(\sigma - KF\tau/Vc)$$

for which the solution with $\tau = 0$ at $z = 0$ is

$$(1) \qquad \tau = T_L - T_R = \frac{Vc\sigma}{KF}\left[1 - \exp\left(-KFz/Vc\right)\right]$$

Let us now consider the function F. Several solutions are available for radial flow from a cylindrical surface in the infinite solid. We require only the gradient of temperature at the radius a. For a steady line-source, coinciding with the axis of the hole, (Carslaw and Jaeger, 1947, p. 221)

$$(2) \qquad F = -4\pi/\mathrm{Ei}(-a^2/4kt), \text{ where } -\mathrm{Ei}(-x) = \int_x^\infty y^{-1}\, e^{-y}\, dy.$$

For small values of a /4kt, this is, approximately, $F = 4\pi/\log(4kt/a^2\gamma)$ with $\gamma = 1.7810$. A better solution for small values of t is derived from the following problem (Carslaw and Jaeger, p. 282): the surface, r = a, of an infinitely long cylinder in an unbounded, uniform medium, initially at zero temperature, is maintained at a constant temperature. From this solution we obtain

$$(3) \qquad F(x) = \frac{8}{\pi}\int_0^\infty \frac{e^{-x\beta^2}\,d\beta}{J^2{}_0(\beta) + Y^2{}_0(\beta)\,\beta}, \text{ with } x = kt/a^2.$$

This integral has been tabulated by Jaeger and Clarke (1942) under the designation, I(0,l,x) for $0 \leqslant x \leqslant 1000$. For kt/a^2 greater than 100, the two solutions differ relatively little.

In figure 2, the calculated distribution of temperature is shown for water entering at a depth of 1000 meters, a = 10 cm, for two rates of flow, 500

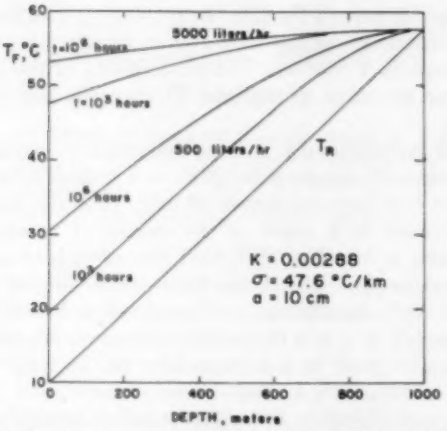

Fig. 2. Computed water temperatures for entrance at 1000 meters. T_R is initial rock temperature; the curves show water temperatures for two rates of flow and for two times after initiation of flow.

and 5000 liters per hour, and in each case for two times, 10^6 and 10^5 hours. The values of conductivity ($K = 0.00288$ cal $cm^{-1}sec^{-1}$ deg^{-1}) and of gradient ($\sigma = 47.6$ °C/km) are based on investigations in the Hungarian plain (Boldizsar, in press). When the flow is small, the water temperature at the surface is close to that of the surface rock (mean annual temperature). When the flow is rapid, the water temperature at the surface differs little from its temperature of entrance. In figure 3, the temperature difference between water and rock at the surface, τ_H, is plotted for the same times and rates of flow, but as function of the depth of entrance H; the thermal constants are as in figure 2.

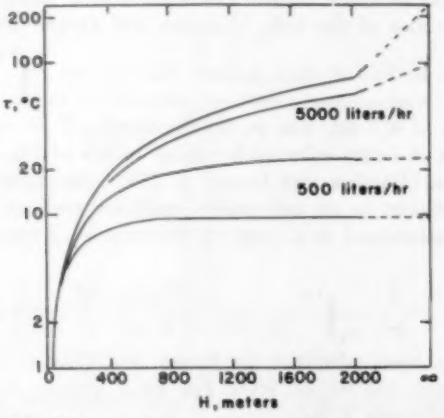

Fig. 3. Computed differences between water and rock temperatures at the surface ($z = H$) as function of depth of entrance, H. The curves are for the same rates of flow and the same times as those of fig. 2.

COMPARISON WITH OBSERVATION

An opportunity for comparison of calculated with observed temperatures in a flowing well is provided by the measurements by Birch (1947) in the L. V. Hart No. 1 well of Colorado Petroleum, Incorporated. The temperatures in this well indicate the entrance of water at a depth of 701 meters, and possibly a smaller inflow at about 500 meters. The observed rate of flow, about eight months after completion of the well, was approximately 1 gal/hour or 63 cm^3/sec. Other required data are as follows:

Radius a, 7.5 cm

Temperature of water at outflow, 16.51°C

Temperature of water at inflow, (2300 feet) 28.40°C

Extrapolated temperature of rock at surface, 14.68°C

Average thermal conductivity of rock, estimated, 0.006 cal/cm.sec.deg

Average thermal difusivity of rock, estimated, 0.01 cm^2/sec

Mean undisturbed gradient in rock, estimated, 19.54°C/km.

We first examine the relationship between $\tau(H)$ and other quantities expressed by (1). As kt/a^2 is approximately 3600, F may be found from the series appropriate for large values of kt/a^2 (Carslaw and Jaeger, p. 282) to be about 1.48. Then KF/Vc is 14.10^{-5} per cm, and for H = 701 meters, exp(–KFH/Vc) is negligible. Thus we have

$$\tau(H) = \sigma/(KF/Vc) \tag{4}$$

Suppose that K is the least well-known of these quantities; on substituting the other values, we find K = 0.005 cal/cm sec deg, a reasonable value (see also Birch, p. 746).

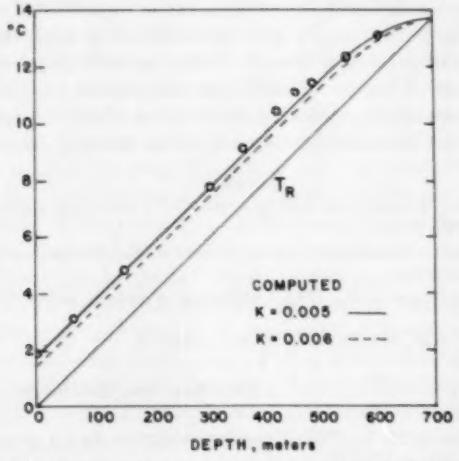

Fig. 4. Comparison of computed and observed temperatures for the L. V. Hart No. 1 well. The circles are the observed temperatures, the curves the computed temperatures for two values of thermal conductivity K. The zero of the temperature scale has been taken at the extrapolated surface rock temperature, 14.68°C, and the undisturbed gradient in the rock is assumed to be 19.54°C/km.

The distribution of temperature in the water can then be calculated from (1); we have

$$\tau = 1.83\,[1 - \exp(-KFz/Vc)] \tag{5}$$

These values are shown in table 1 and plotted in figure 4. The calculated temperature difference has been added to the supposed undisturbed rock temperature to give a calculated curve for comparison with the observed temperatures. Agreement is good except near 500 meters, where the temperatures suggest a possible small influx of water. The calculated curve for $K = 0.006$ is also shown; the agreement is less good than for $K = 0.005$. It should however be kept in mind that the parameter which determines these curves is essentially K/V; thus an error in V produces a proportional effect on K, as found in this way. Since the flow for this well is known only for a single date, it may be considered that the agreement is as good as can be expected. In addition, the curve for undisturbed rock temperature is based on extrapolation from greater depths, so that $\tau(H)$ is uncertain. The extrapolated temperature of rock at the surface is appreciably higher than the mean annual temperature, though measured values showing equally large differences are not unknown (Van Orstrand, 1939).

TABLE 1

Computed Difference of Temperature Between Water and Undisturbed Rock

z, meters	0	50	100	200	400	600	701
$[1 - \exp(-KFz/Vc)]$	0	0.48	0.73	0.93	0.995	0.999	0.9999
τ, °C	0	0.85	1.33	1.70	1.82	1.83	1.83

In principle, if the flow were well observed and found to be steady, and if the other assumptions of the theory were approximately satisfied, observations in a flowing well could be used, in the way indicated above, to give a mean conductivity of the rocks, and in combination with the mean thermal gradient, a value of local heat flow. It is also possible to generalize this treatment for the case of layers of different conductivity, for inflow at several levels and at temperatures differing from those of the corresponding rocks, and for gases which are cooled by expansion on entering the well.

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$$\int_0^\infty \frac{e^{-xu^2}}{J^2_0(u) + Y^2_0(u)} \frac{du}{u}$$

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ESSAY REVIEWS

Glacial and Pleistocene Geology; by RICHARD FOSTER FLINT. P. xiii + 553; 134 figs., 5 pls., 51 tables. New York, 1957 (John Wiley & Sons, Inc., $12.50).—Since *Glacial Geology and the Pleistocene Epoch* appeared ten years ago as the first and sorely needed North American textbook on the subject, Pleistocene research has accelerated at such a rate that only those truly interested could expect to keep themselves properly informed on their specialties. Geologists the world over now are indebted for this current up-to-date summary of glacial and Pleistocene geology. The greatest advances have been in related fields. The physics of and meteorology of glaciers (glaciology), dating by radiocarbon and by measurements of paleotemperatures (geochemistry), significance of paleosols (paleopedology), interpretation of cores of sea floor sediments (marine stratigraphy), importance of frozen ground phenomena (cryopedology), pollen analysis and interpretation of lake sediments (palynology)—the results of recent research in all these fields are here amplified for the professional. This then is in part a compilation of a decade of Pleistocene research.

Much of the compactly written basic or standard glacial geology of the earlier volume also is retained. Some chapters are completely reorganized and rewritten, others revised and brought up to date. The first 17 are glacial geology (301 p.); the last 10 are Pleistocene geology (207 p.) dealing with Pleistocene geography and stratigraphy of North America, of Europe, of areas outside of North America and Europe, with deep-sea floor stratigraphy, with Pleistocene fossils, and with the causes of glaciation. The quite new chapters, although containing some material familiar to readers of the first book, are Frozen-Ground Phenomena (12 p.), Weathering and Soils; Stream Regimens (15 p.), Fluctuation of Lakes in Nonglaciated Regions (18 p.), Crustal Warping (18 p.), and Stratigraphy of Deep-Sea Floors (7 p.). In the chapter on Chronology, the explanation of radiocarbon dating and its shortcomings is most lucid. Most correlations, some made boldly and others too guardedly, are offered tentatively only in table form. Of the 49 photographs, 15 are well reproduced aerial views. The many maps either are more legibly redrawn or are new. The list of 866 references (up to 1957) is admittedly an attempt only to support the text. The whole book is printed on glossy paper, a fact undoubtedly accounting in part for the regrettably high cost.

The excellently written chapters on elements of glaciology emphasize the need for personal firsthand knowledge of glaciers before significant analyses of their many effects can be made. These brief chapters (2 & 3, 35 p.) do not satisfy the need for a single publication from a North American writer on the principles of glaciology and physics of glacier ice, although several reviews of glaciology do exist in other books, e.g.: Part I of Charlesworth's *The Quaternary Era*, 1957, and Part I of Lliboutry's *Nieves y Glaciares de Chile*, 1956. As much of the information derived from glaciological studies of the Greenland ice sheet as is now available is integrated into a smooth account of the regimen of the ice sheet, and pertinent data are listed for comparison with the former ice sheets of North America and Europe. In the chapter on Frozen-Ground Phenomena, the awkward but more easily understood (by English-speaking people) words, such as frost wedging, solifluction, solifluction

deposits, and permafrost unfortunately are still favored over the etymologically correct, more universally useful, and now accepted (especially by Europeans) terms introduced by Bryan in 1946.

Controversial matters are animated with information published during the past ten years. Evidence supports the idea of two types of cirques, those carved by frost wedging and glacial erosion (true glacial cirques), and those incipient cirques excavated by nivation at sites of firn banks indicating merely a frost climate. The existence of mature forests at the edge of much of the North American ice sheet is now accepted. The request to abandon the no longer adequate concept of a "Tertiary period" and a "Quaternary period" again is suggested as most desirable, and the petition to use the terms recent and postglacial only informally and locally is solicited anew. The latter position is still (see THIS JOURNAL, vol. 255, p. 385-393) strongly opposed by some members of the U. S. Geological Survey. The problem of interpretation of two tills in New England is entertained once more, with the conclusion that the younger till probably represents "the classical Wisconsin" and the older may antedate the Wisconsin.

One cautionary change that may shock the reader searching for a correlation of a certain favorite moraine, loess, or soil with one of the standard Wisconsin subdivisions is that any such correlation purposefully has been avoided. The substages of the Wisconsin, formerly supported, are mentioned but once and then only to exemplify subdivisions that must "necessarily be local," and the appropriateness of correlating other drift sheets with these substages is questioned. Instead, the phrase "classical Wisconsin drift" with all its implications is used throughout. The closest to a correlation offered is the casual, almost second-thought, listing of central North America equivalents of "Post-Sangamon" events (Table 23-B) in a chapter on the Stratigraphy of Europe. This conservative approach otherwise is the thesis.

The final chapter on the causes of glaciation is exceptionally well done, especially in the logical presentation of the solar-topographic concept. The implications of recently acquired meteorological data are employed to the full in the build-up of a seemingly watertight case. Variations in the constituents of or turbidity of the atmosphere, crustal or polar displacements, or changes in the axis or orbit of the Earth become here either insignificant or solely ancillary.

SIDNEY E. WHITE

Structural Geology; by L. U. DE SITTER. P. 552. New York, 1956 (McGraw-Hill Publishing Co., $11.00).—A refreshingly new viewpoint has been introduced into the structural field by Professor de Sitter. He is at the University of Leiden in Holland and, as one might expect, his approach differs somewhat from that of the conventional texts in common use in the English-speaking world. Not only are many of his examples new, selected from the furthest corners of the earth, but telling his tale in the first person injects a familiar note, a you-and-me conversational style that makes pleasant reading. This adds an appreciable number of pages to the text, but readers will have to judge for themselves whether the increased volume of words is really useful in the long run.

The book is divided into three parts: I. *Theoretical Structural Geology* (physical properties of rocks, laboratory experiments, rock fracturing and distortion), II. *Comparative Structural Geology* (systematic account of faulting, folding, and primary sedimentary features), and III. *Geotectonics* (magmatic phases in orogenesis, different types of fold belts, island arcs, shields, basins, theories of orogenesis).

Very clearly, we have here a unique work filling a serious gap in geological literature. It will not compete with any other publication, because there is nothing else quite like it. Having used it for one season as a senior classbook, this reviewer found that the students were unanimously enthusiastic about it. This should be recommendation enough.

The work was not planned as a junior text, simple, dogmatic, and empirical. Rather it is set at a senior or graduate level, in the form of a series of descriptive conversations on numerous aspects of structure and geotectonics. Its most important contribution, as we see it, is its clear account of selected examples of *real* mountain chains (e.g. the Alps, Pyrenees, High Atlas), all of which differ fundamentally from one another. It quickly passes on from the arid catalogs of terminological jargon that clutter structural literature and plunges into the exciting realm of actual observation.

Professor de Sitter adopts an historical approach in many cases, with excellent bibliographic references (about 500), mostly to English-language papers, but not omitting the foreign where there is no English version. Occasional reference to stratigraphic series in the European time-scale may sometimes confuse readers in North America; it seems a pity to speak of "Lias" where "Lower Jurassic" would have been equally valid and much more readily understood. One may be a little irritated by the constant reference to local examples (in the Pyrenees, for instance) in order to explain such prosaic topics, say, as boudinage or slaty cleavage. These things are so universal that they hardly require specific examples to establish their reality.

Some curious ommissions are noted: although there is a chapter on "Structural petrology, cleavage, and schistosity," there is no reference to foliation or lineation, except as may be read into the discussion on preferred orientation patterns. In fact, de Sitter gives very little attention to structures in metamorphic or igneous rocks, and in his first lines he applies the axe rather sharply: "Structural geology is concerned with analyzing the deformation of sedimentary strata." The strain ellipsoid is given short shrift; its limitations are fairly noted, after which it is written off as a dead loss. This reviewer feels that it still has great virtues, providing it is not used as a sort of divining rod.

In the last decade or so, much nonsense has been written and quite a smoke screen of confusion has been set up on the subject of flysch-molasse facies. For those in trouble, please read de Sitter's Chapter 20 for a concise and clear statement of the facts. These are extraordinarily valuable terms for the tectonicist, but if misused will only end in his disaster.

On the somewhat vexed topic of geosynclines, de Sitter follows the ultraconservative line, pre-Stille, pre-Kay. "Geosynclines," pure and simple, thus refer only to orthogeosynclines; all parageosynclines are "intercratonic basins"

(more accurately, intra-cratonic). Many readers will like this treatment, for it enables the sedimentologist to speak of "geosynclinal" versus "epeirogenic" facies without modification. However, those of us who find great value in the Kay terms will regret the move.

The standard of printing, paper, binding, and so on are excellent, as one would expect from McGraw-Hill. Printers' errors have been kept to a minimum: may heaven forgive "peneplenization" (on page 9). The illustrations, mainly line, represent a really magnificent effort. The labor is amply justified. In a few cases, the reduction has been excessive and legibility impaired; such pennywise blunders by publishers could even be overcome by another dollar on the price without scaring off potential customers. The present trend in book prices is not out of line with the contemporary cost-of-living, tragic as it is.

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VOL. 256 No. 4, 1958

American Journal of Science

APRIL 1958

CONTENTS

Properties of Water, Part II. The Equation of State of Water and Sea Water at Low Temperatures and Pressures ..... *Carl Eckart* 225

Interaction of Basement and Mantle During Folding Near Rutland, Vermont ..... *W. F. Brace* 241

Idiomorphic Zircon, Key to Volcanism in the Lower Tertiary Sands of Central Texas ..... *Dean L. Callender and Robert L. Folk* 257

Narrow Linear Drumlins Near Velva, North Dakota *Richard W. Lemke* 270

Periglacial Loess, I. Morphological Properties *J. F. Gerard Millette and Howard W. Higbee* 284

The Distribution of Temperature in Flowing Wells ..... *T. Boldizsar* 294

ESSAY REVIEWS

Glacial and Pleistocene Geology; by Richard Foster Flint ..... *Sidney E. White* 299

Structural Geology; by L. U. de Sitter ..... *Rhodes W. Fairbridge* 300

PUBLICATIONS RECENTLY RECEIVED 302